# THE LAST GLACIATION

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# THE LAST GLACIATION

## With Special Reference to the Ice Retreat in Northeastern North America

BY
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# PREFACE

Five years ago Dr. Antevs' first contribution to American glacial studies appeared in this Research Series under the title "The Recession of the Last Ice Sheet in New England." As the first successful working out of an exact glacial chronology on this side of the Atlantic, it was unique. Nature's own record of 4300 years in New England was accurately deciphered by a method already in use in Sweden but until then not well known in America. Moreover this was the first pronouncement of the principle that, if we are to correlate the ice retreat in America with that in Europe or elsewhere, we must first discover and measure the larger variations in rate of recession, due to slow, great oscillations of climate, and then, comparing these long fluctuations see whether or not they correspond in the two cases —not trusting a direct year-for-year comparison of varve graphs for localities on opposite sides of the ocean or north and south of the equator.

Those who have followed Dr. Antevs' investigations since 1921, as they are recorded in several memoirs and shorter papers, know that he has worked back and forth over New England, has traversed the St. Lawrence lowland and the upper Great Lakes region, and has gone far northward towards Hudson Bay, not stopping where railways and highways end, but proceeding on by canoe through the wilderness, in his quest for more exposures of the precious laminated clays. Also, he has spent a summer in a re-study of the mysterious climatic history of the Great Basin region of Nevada, Utah, and California. So, he has come to be as intimately acquainted with North America as with northern Europe.

The patient, accurate measuring of clay layers, or *varves*, the plotting and matching of graphs, and the cautious interpreting of these as yearly records of climatic change has been a vital part, yet only a small part, of his work. Equally at home in physiog-

raphy, Pleistocene botany and zoölogy, and eager in his study of climates of the past, he has never lost sight of the large geologic and geographic problems, such as the registration of changes of temperature and of rainfall by sediments, by tree rings, and by migrations of plants and animals, the nature and cause of crustal warping during the withdrawal of the ice, the emergence and submergence of the coast and of the interior lowlands as a result both of crustal movements and of the exchange of water between ocean and glacier.

In so far as it presents evidence not hitherto given out, this book is another step forward in the attack on fundamental questions of the Ice Age; but more than that, it is a deliberate pause, on the part of this versatile and thoroughgoing scientist, to consider what other investigators the world over have been finding and thinking. The extraordinary collection of references and the compact summaries of opinion show the author's intention to hear all sides before he draws conclusions. His knowledge of the literature of American Pleistocene history and his regard for what others have done in this field are particularly noteworthy.

It is certain that not all who read and study Dr. Antevs' book will concur with him in his views. *Were* the glaciers in the various parts of the world synchronous? But since we cannot profitably discuss rival theories as to the cause of the Ice Age nor settle many vexed questions of Pleistocene history until we have a fair answer to this fundamental one, all will welcome this carefully considered statement of the views of a man who has been searching diligently for the truth, over the whole field of Quaternary geology and geography, and who now puts his facts, his elaborate measurements, and his thoughts beside the facts and opinions of the specialists of seven continents.

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# INTRODUCTION

The Quaternary is one of those cold and climatically unsettled periods in the earth's history which at long intervals have broken the warm and more equable climatic conditions and have thus greatly influenced the evolution of plants and animals. As the latest epoch it furnishes criteria for geological processes in general as well as for the earlier ice ages and their rôle for physiography and life. It also gives the clue to the existing state of things on the earth's surface; to its physical features, and to the migrations and distribution of living plants and animals. In the hope of shedding light on various problems an attempt is here made to treat and reconstruct from scattered material some important physical conditions at the climax of the last glaciation with attention especially to the recession of the last ice sheet in the peripheral belt of the glaciated area of eastern North America, in the frontier regions between the eastern United States and Canada, and in northern Ontario and Quebec. Since this book is the last of a series (on the late-glacial epoch of North America), the general problems here discussed have been selected so as to supplement those previously treated.

The material here published was for the greater part collected in 1922, 1924, and 1925, and some of it in 1921 and 1923. All of the material was worked up at Harvard University. The field work of 1921 and 1923 was supported by the National Research Council of Washington and the Geological Survey of Canada, respectively. That of 1922 was made possible by a grant from the University of Stockholm obtained through the courtesy of Baron Gerard De Geer. Since July, 1924, the studies both in the field and in the office have been supported by the income from the Shaler Memorial Fund, of Harvard University, for which the writer wishes to convey his sincere thanks to the Division of Geology.

The late Professor J. B. Woodworth, Professor Reginald A. Daly, Professor Charles Palache, Professor Robert DeC. Ward,

and Mr. Robert W. Sayles of Harvard have given much good advice and help and shown the greatest courtesy and interest. Professor A. P. Coleman, the late Professor James F. Kemp, Professor William H. Hobbs, Mr. Frank Leverett, Professor C. W. Brown, Dr. W. H. Collins, Mr. W. A. Johnston, Dr. John L. Rich, Professor H. F. Cleland, and the late Dr. H. H. Robinson have given valuable information and advice. Professor J. W. Goldthwait, in whose agreeable company part of the field work in 1925 was done, has aided, as always, in many ways.

Finally, the writer is greatly indebted to the American Geographical Society for publishing the work.

# CHAPTER I

## WERE THE PLEISTOCENE GLACIATIONS IN THE VARIOUS PARTS OF THE WORLD SYNCHRONOUS?

The question of contemporaneity of former glaciers in various parts of the world is almost as old as the glacial theory itself. Within a few years after Agassiz had drawn the bold picture of a former ice sheet overspreading the Alps, this was duplicated by the discoveries of Lyell in England and Edward Hitchcock in America, although the real nature of the "drift agency" was long debated. One by one, the signs of old glaciers have been seen and recognized all over the world, while their survivors, the living glaciers, have been located and studied from the poles to the equator. Do these ancient ice sheets and mountain glaciers register one simultaneous chapter of climate?

Especially during the last fifteen or twenty years, information on glaciation has come in more rapidly and has been fuller and more trustworthy because explorers have been better equipped for scientific work. The picture of Pleistocene geography is thus coming to take clearer form. With this advance in knowledge of the distribution of former glaciers has come a growing conviction that everywhere the glaciation was multiple, consisting of three or four successive epochs, as it has long been known to have been in the case of the American and European ice sheets. Accordingly we may now feel justified in forming an opinion as to the degree to which glaciation was synchronous. The more complete proof or disproof of the question, by careful comparison of oscillations in ice recession as registered by successive moraines whose dates and ages may be accurately determined by measurement of well connected series of varved clays, is one of the ultimate aims of studies like this.

The correct answer to this question will determine the fate of more than one theory of the cause of the Ice Age, particularly such as are based upon changes in the earth's position relative to the sun. Questions of a secondary character, yet of just as much interest to geologists and geographers, such as the amount of fall and rise of sea level as ice sheets grew and shrank, and the migrations to and fro across the continents of animals and plants as glaciers waxed and waned—these, likewise, remain unsolved until we know to what extent climatic changes were synchronous.

## Northern Hemisphere

### NORTH AMERICA

It has been a widely accepted view (originating with Tyrrell, 1896, 1898, 1914) that during the Pleistocene there was an eastward migration of the centers of glaciation in North America. This was first suggested by studies on striae, etc. in the Hudson Bay region. The Cordilleran, Keewatin, Patrician, and Labradorean ice sheets are supposed to have arisen, reached their greatest extent, dwindled, and perhaps nearly disappeared more or less in succession, while the center of glaciation steadily moved eastward.

However, this view lacks support from facts in the intermediate and peripheral belts of the glaciated area. What seems to be the very earliest Pleistocene glaciation affected New Jersey and Pennsylvania and possibly Long Island (Veatch, 1906, p. 34; Fuller, 1914, p. 85) as well as Nebraska. Also other early glaciations have left records in Long Island (Veatch, 1906, p. 33; Fuller, 1914, pp. 92, 132, 150, 157). The moraines marking the extreme limit of the last, or Wisconsin glaciation extend in a continuous series from Cape Cod on the Atlantic all the way to Alberta. The most recent traces of extensive glaciation in the extreme West are quite as fresh as those in the East (Leverett, 1917; see Antevs, 1925a, pp. 67, 74). Thus each of the glaciations has affected both the East and the West, and the successive ice sheets and glaciers have covered practically the same areas.

The last glaciation may have reached its climax at different centers, however, at somewhat different times (Leverett, 1916; 1919, pp. 15, 53; 1926, p. 118; and communication in letter). The Labrador field was the first one to attain its maximum. The Patrician sheet or subdivision culminated at the time of the formation of the Kettle morainic system of the Green Bay ice lobe and its correlatives. The Keewatin ice reached its greatest extent contemporaneously with the formation of the Port Huron morainic system. Still, it appears from Tyrrell's studies that after the ice had melted away from the district south of Hudson Bay, there came a late advance from Labrador.

### EUROPE

In Europe an eastward shifting of the Pleistocene glaciations has recently been discussed (Limanowski, 1922; also Kulczyński, 1924, pp. 181–185; Kozlovskii, 1924, p. 115). Of the assumed four glaciations the first is believed to have occupied only Great Britain and Scandinavia, the second similarly only western and central Europe, the third western and central Europe as far as western Russia, and the fourth the area from the North Sea to eastern Russia. The greatest extent was reached in western Europe during the second glacial epoch, in central Europe during the third, and in eastern Europe during the fourth. The theory, according to Kulczyński, is in accord with phytogeographical conditions. However, G. F. Mirčink holds that Russia suffered four Pleistocene glaciations; and there are various interglacial deposits in eastern Europe attributed to the second interglacial interval (see Dokhturovskii, 1925, pp. 81, 104, Tab. 1 and 3; in Tab. 3 deposits believed to date from the second interglacial have erroneously been placed in the third; Schultz, 1927). Northeastern Russia was glaciated two or three times (Grigoryev, 1925, p. 119), and Caucasus probably three or four times (von Reinhard, 1925).

On the whole the glaciations in the various parts of Europe are believed to have been synchronous.

## NORTH AMERICA AND EUROPE

The last glaciations in North America and in Europe according to all indications were contemporaneous, and the waning of the ice sheets may have been parallel events corresponding at least in their large features (see p. 163). Also the earlier Pleistocene glaciations in the two continents probably were synchronous, judging from the corresponding degree of weathering and valley sculpture of the drift deposits (Leverett, 1910; Leverett quoted by Osborn & Reeds, 1922).

## ASIA

In Asia, likewise, glaciation was widespread and recurrent.

The extent of the glaciations in Siberia is very unsatisfactorily known, but fairly large areas were ice-covered not only in the mountains but also in the northern plains. At the eastern foot of the Ural Mountains terminal moraines reach down to the latitude of 62¼° N. and from there extend in north-northeasterly direction to the southeastern part of the Taimyr Peninsula (Obruchev, 1926, p. 388, Pl. 10). Lenin Land and the New Siberian Islands were heavily glaciated. The ice in the latter islands according to Obruchev was possibly connected with the Taimyr ice. In northeastern Siberia east of the Lena River all the high mountains, viz. the Tomuzkhaya, Ulakhan Chistai, Tai-Khayakhtakh, Charaulachsk, Verkoyansk, Kolimsk, and Anadir ranges, and the mountains at Cape Deshnev on Bering Strait were glaciated (Obruchev, Pl. 10). According to Sevastyanov (1910) even the whole area drained by the Yana, Indigirka, and Kolyma Rivers was occupied by a continuous ice cap. In Molchanov's opinion (1926) the whole region from the Urals to the Pacific north of the 61st or 62nd parallel was glaciated. West of the Sea of Okhotsk, the Aldansk and the Stanovoi Mountains carried glaciers. Northeast of Lake Baikal, in the mountainous regions of the upper courses of the Olekma and Vitim Rivers, glaciers had extensive distribution and were thick in some places, covering valleys and mountains. South of Lake Baikal, the Kentei

Mountains show evidence of glaciation. West of Lake Baikal, mountain glaciation was extensive (see Hausen, 1925, p. 1120; Obruchev, 1926, p. 390, Pl. 10). The East Sayansk Mountains were covered by a large ice cap sending lobes down the valleys. The West Sayansk Mountains were also glaciated, as were the Kuznetsk Alatau. Tannu Ola locally carried small glaciers. The Altai, even now glaciated, during the Ice Age were heavily ice-covered. Adjacent mountains in northwestern Mongolia and in Dsungaria show traces of Pleistocene glaciers (see Obruchev, 1926, p. 391; also S. V. Obruchev, 1927).

In southern Kamchatka there was a great network of glaciers, which on the east coast reached the sea (Komarov, 1912). In Japan the Hida Mountains (36° N.) held small hanging glaciers at altitudes of 2500 to 2600 m. (8200 to 8500 feet) (Simatomai, 1914; Oseki, 1914; Yamasaki, 1922). In central China, on Taipai-shan (34° 20′ N., 107° 30′ E.; 3550 m., 11,650 feet), the highest summit of the Tsinling-shan, the former existence of glaciers is shown by cirques at 3440 m. (11,290 feet) elevation (Limpricht in Olbricht, 1923, p. 727). In southwestern China, in the Yunling Mountains (30° N., 103° E.)—the only Chinese mountains that at present rise above the snow line, which on the 30th parallel lies at about 5500 m. (18,050 feet)—the snow line during the Pleistocene was about 1400 m. (4600 feet) lower than the modern as shown by cirques, U-shaped valleys, etc. Still farther south the mountains between the Saluen and the Mekong (about 27° N., 99° E.) show signs of strong former glaciation (Handel-Mazzetti, 1923).

In central Asia, from the Himalayas and northward, all the highest mountains seem to have been quite extensively glaciated (see Geikie, 1895, Pl. 13, p. 691). In the Pamir plateau, the Alai Mountains and adjacent plateaus and mountain ranges the glaciation was so extensive as to attain the character of continental ice (von Ficker's review of Klebelsberg, 1922). In Turkestan the glaciers, without forming a *mer de glace*, descended on an average 1000 m. (3300 feet) lower than now (D. I. Mushketov; see J. V. Fuller, 1925, p. 662).

The Caucasus was extensively glaciated (von Reinhard, 1914; von Stahl, 1923, p. 43, Pl. 2). Various mountains in Armenia carried glaciers at lower levels than today (Oswald, 1912, p. 19).

In several areas, as Kamchatka, northeasternmost Siberia, the highland of the Olekma and the Vitim Rivers, Mongolia, Turkestan, etc., records of two separate glaciations have been obtained, the former of which was more extensive;[1] and it is to this one that the data just given have reference. The last glaciation was less extensive. In Shansi, China, three Pleistocene glaciations appear to be recorded by three different beds of loess (Solger, 1914; Olbricht, 1923). In the western Himalayas four glacial epochs have been distinguished and tentatively correlated with the three latest ones and the first post-Würmian stage in the Alps (Dainelli, 1922; see also Loewe, 1924).

North of the northern end of the Ural Mountains birch occurs between two beds of till far north of the present range of its occurrence (Backlund, 1925, p. 502).

On Lake Nelgato, 70½° N., west of the mouth of the Yenisei and 400 km. (250 miles) north of the present tree line, remains of mammoth were found in clay together with twigs and leaves of birch (*Betula nana*), willow (*Salix glauca* and *S. herbacea*), and wood and roots of tamarack (*Larix*) (Schmidt, 1872, p. 33; cf. Walther, 1919–22, p. 453). The mammoth bed is underlain by marine clay with arctic shells and overlain by beds of clay in alternation with layers of vegetation containing mosses, twigs of willows, and leaves. The mammoth is evidently either interglacial or postglacial.

In Taimyr Peninsula fossil wood, especially of Larix, is frequent up to the latitude of 74½° (Backlund, 1925, p. 502). The modern tree line lies at the 72nd parallel. The wood, which consists partly of quite large logs, rests on permanently frozen ground and is locally overlain by ground ice. It may be of interglacial age. The northern limit of the wood determines the limit of the summer migrations of the nomads.

[1] In the Altai (Fickeler, 1925, 1925a, p. 633) and in the Romanov Mountains in western Turkestan (Klebelsberg quoted by Loewe, 1924, p. 53) reliable evidence of but one glaciation has thus far been found.

In the New Siberian Islands the Quaternary deposits present the following strata according to K. Volossovich (see Brooks, 1925, p. 79; Obruchev, 1926, p. 392):

Topmost, marine clay, deposited in a climate somewhat warmer than the present.

Clay with dwarf birch, arctic willow, and bones of musk ox, horse, and later mammoth.

Fossil ice.

Fine clay, with remains of alder and white birch and bones of mammoth and rhinoceros, changing downward into sandy clay with remains of meadow vegetation and shrubs.

Fossil ice.

Before the deposition of the top clay submergence took place; and after its formation emergence followed, inaugurating the existing conditions. The fossil ice is considered by the Russian geologists to be the buried remains of the Pleistocene ice caps and glaciers. Thus the section may record two glaciations, an interglacial epoch, and postglacial time. The mammoth remains in these islands, according to Baron Eduard von Toll (see Walther, 1919–22, p. 454; Köppen and Wegener, 1924, p. 190), are most frequently associated with a forest vegetation of willow (*Salix* forms), birch (*Betula nana* and *B. alba*), and alder (*Alnus fruticosa*) and thus, judging from the above profile, are for the most part of interglacial age. The trees enumerated are now, according to Köppen, not to be found until five to six degrees of latitude south and indicate a climate comparable to that of the present interior of eastern Siberia. There is much other evidence of higher temperature in northern Siberia during the Quaternary (see Obruchev, 1926, pp. 383–386, 392–397).

The well-preserved mammoth found on the Beresówska River in northeastern Siberia had in its stomach and mouth plants of a meadow flora such as characterizes the region at the present time (Salensky, 1904) and is surely not of glacial age.

Thus there are from many parts of Asia records of two to four Pleistocene glaciations and from northern Siberia evidence of an interglacial stage with essentially higher temperature than that now prevailing. These conditions, together with the general

distribution of the former ice sheets and glaciers on the northern hemisphere and especially the proximity of some of the glaciated mountains in Asia to the European ice sheet, convincingly show that the glacial and interglacial epochs were essentially synchronous in all the northern hemisphere. The conditions as far as known do not support the frequently advocated view that Siberia and also Japan (Yokoyama, 1911, 1920, 1922–23; Yokoyama's view is challenged by Yabe, 1922, and by Yamasaki, 1922) during the greater part of the Quaternary time, even when northern Europe and North America were covered by huge ice sheets, enjoyed higher temperature than at present and consequently that the north pole was situated far from its present position and was migrating over Greenland. Against this view also may be set the evidence of the plant and animal remains in the interglacial beds at Toronto, which indicate higher temperature than that in modern time (Coleman, 1913a, p. 18).

## Southern Hemisphere

### SOUTH AMERICA

In South America signs of Pleistocene glaciations much more extensive than the present one have been traced down the whole length of the continent, from northern Colombia to Cape Horn, in spite of the fact that these steep cone-shaped mountains are not favorable for the development of glaciers. The evidence shows certainly two and perhaps three Pleistocene glaciations (Meyer, 1904, 1907; Hauthal, 1911, p. 175; Sievers, 1911, p. 200; Bowman, 1916, p. 206; McLaughlin, 1924; Klute, 1925; Mortensen, 1927), but their extent is only partly known; and it is not in all cases clear to which glaciation the statements of authorities refer. However, the oldest one was the most extensive.

In northern Colombia, in the Sierra Nevada de Santa Marta (11° N.), the glaciers extended at least 1200 m. (3950 feet) lower than at present, or down to levels of 3400 to 3500 m. (11,150 to 11,500 feet) and in Venezuela, in the Sierra Nevada de Mérida

(8° N.), 700 m. (2300 feet) lower, or to an altitude of 3800 m. (12,500 feet) (Sievers, 1908, pp. 274, 279). In Ecuador, near the equator, the glaciers in geologically recent time reached 800 to 900 m. (2600 to 3000 feet) lower than today, that is to altitudes of 3700 to 3800 m. (12,100 to 12,500 feet) (Meyer, 1907).

In northern Peru the lower limit of the glaciers during the last glaciation was at 3900 m. (12,800 feet) altitude (Sievers, 1914, p. 253); in central Peru (10° to 11° S.) at 4000 m. (13,100 feet); in southern Peru at 11,200 feet (3400 m.) (Bowman, 1916, p. 211); in eastern Bolivia (16° to 21° S.) at 4100 m. (13,500 feet); and in northeastern Argentina (24° S.) at 4500 m. (14,800 feet), that is 800 to 1000 m. (2600 to 3300 feet) below the present limit. However, there was at least one preceding glaciation, during which the glaciers seem to have descended from 1000 to 1200 m. (3300 to 3900 feet) and the snow line 400 to 500 m. (1300 to 1600 feet) still lower than during the last one (Hauthal, 1911, p. 181; Bowman, 1916, pp. 204, 207, 211).

The Pleistocene glacial limit from 11° N. to 11° S. according to Sievers (1911, p. 194) lies at practically the same level, south of which it rises rapidly and is divided in two lines, one on each side of the mountain range. In Bolivia and farther south the lower limit lies at very different levels.

In the southern part of the continent the glaciers during their greatest extent, that is during the first glaciation, descended from the high mountains down on the plains and reached their greatest expansion towards the south (Nordenskjöld, 1909, pp. 109–112). In Patagonia, however, the ice did not reach the sea, and the greater part of the country was not glaciated. Only south of the 52nd parallel did the ice extend to the eastern coast of the continent. Thus no typical ice sheet was developed on the southern hemisphere outside the Antarctic (Nordenskjöld, 1909, p. 110). The extent of the last glaciation is not well known.

The Falkland Islands were not glaciated (J. G. Andersson, 1906, p. 101).

### AUSTRALIA AND NEW ZEALAND

In the Australian continent the only evidence of Pleistocene glaciation is to be found on the Kosciusko plateau in southeastern New South Wales, which has an elevation of about 7000 feet (2130 m.) (David, 1908; 1923, p. 130; Howchin, 1918, p. 492). Records of two or three glacial epochs have been found. During the first and most extensive epoch the whole plateau was capped by glaciers, which came down to within 4500 feet (1375 m.) of sea level and covered an area of 80 to 100 square miles (200 to 260 square km.). During the last glaciation the ice extended down to the level of 6150 feet (1875 m.).

In Tasmania there are likewise records of at least two if not three glacial invasions (Lewis, 1921; 1923, p. 32; 1924, p. 38; G. Taylor, 1921; David, 1923, pp. 115, 130; Clemes, 1924, p. 69; cf. Benson, 1916, p. 35). During the earliest one, which was by far the most extensive, a more or less continuous ice cap covered the whole central plateau, or fully one-third of the island; and glaciers descended the valleys and at places reached the sea. In the National Park, moraines were formed at altitudes of 2500 to 2800 feet (760 to 850 m.). The ice, probably 1500 to 2000 feet (450 to 600 m.) thick, covered valleys and mountains except the highest peaks that formed nunataks. During the last less severe glaciation lakes were formed by ice action in the National Park at altitudes of 3200 to 3500 feet (975 to 1065 m.).

In the South Island of New Zealand, where now only a few glaciers are to be found in the mountains of the interior, the glaciation in Pleistocene time was considerable (Marshall, 1910; 1912, pp. 31, 49, Fig. 13; Speight, 1921, p. 304; cf. Park, 1910). However, it was limited to the highlands and consisted of separate glaciers rather than a continuous ice cap. In the North Island there is, to be sure, no certain evidence of glacial action; but this lack might be due to obliteration by postglacial volcanic eruptions (Moore, 1917; Benson, 1921, p. 114).

### SOUTH AMERICA AND AUSTRALASIA

Thus in both South America and Australia evidence of two or three Pleistocene glaciations has been found. In both regions the first one was most extensive. The degree of decomposition of the drift and the extent of other postglacial alterations seem to correspond. All this suggests that the glaciations in the two continents were synchronous. They cannot have alternated since this would postulate very improbable migrations of the poles, contrary also to the evidence from the northern hemisphere. That the south pole during the Pleistocene cannot have been very far from its present position is also evidenced by former more extensive glaciations or cooler climate not only in South America and Australia lying on the opposite sides, but all round it. The mountainous island of South Georgia (54° S., 37° W.) presents traces of much heavier glaciation than today; and the Kerguelen Island (49° S., 70° E.), on which now névé collects only on the higher parts and from there descends as glaciers to sea level, not very long ago was entirely or almost entirely ice-covered (Nordenskjöld, 1909, pp. 131–136; Krenkel, 1925, p. 393). South Africa had a cooler climate (see p. 43). However, Köppen in a recent paper (1927) declares his belief in a considerable migration of the south pole.

### THE ANTARCTIC

However extensive the present glaciation of the Antarctic, it was once still greater. Indications of this are to be found everywhere inside the polar circle. At Gaussberg (67° S., 90° E.) the ice was at least 240 m. (785 feet), in Victoria Land 120 to 150 m. (400 to 500 feet), and at Robertson Bay (71 ¾° S., 170° E.) 300 m. (1000 feet) thicker than at present (von Drygalski, 1919, p. 26; Brückner, 1913, p. 276; David, 1914, p. 622; Wright and Priestley, 1922, p. 438). The Ross Barrier has lost at least 800 feet of ice on the surface, and its edge has retired southward 200 miles (320 km.), in places even 20 to 30 miles (30 to 50 km.) since 1840 (David, 1914, p. 622). This is just in the area where, on account

of its proximity to moisture-bearing currents, one would expect the glaciation to be at its maximum now. Farther in, the Beardmore Glacier (84° S., 171° E.) was once at least 2000 feet (600 m.) thicker. Probably about 1000 feet (300 m.) of ice has disappeared from the ice sheet since its maximum (David, 1914, p. 622). Only locally and especially at the lowest latitudes does the ice have its maximum extent at present.

The date of the greatest extent of the Antarctic ice sheet is uncertain, views differing according to opinions as to the chief cause of glaciation and as to the mode in which the ice is nourished. At present thaws occur only locally and at infrequent intervals even in the middle of a normal summer. All precipitation is in solid form, and rain is almost unknown (Wright and Priestley, 1922, pp. 24, 280). Obviously, then, either a greater snowfall or a more effective accumulation of the fallen snow is required to explain the former heavier glaciation.

Various writers like Scott (1905, Vol. 2, p. 425) and Brückner (1913, p. 279; see also Wright and Priestley, 1922, p. 455; Hess, 1924, p. 330) think that the greater glaciation occurred during a period of higher temperature than the present. The argument is based on the facts that snowfall in the peripheral regions is greater in the warmer months, that accumulation and consolidation of snow is most effective at temperatures near the melting point, and that islands such as Bouvet Island (55° S., 5° E.) show intense glaciation in spite of their much higher mean annual temperature. In Snow Hill Island, for instance, the ice grew more during the warm season than during the cold (Nordenskjöld, 1911, p. 139).

On the other hand, since snowfall over the interior of the ice is of first importance, since a strong permanent anticyclone existed as at present over the ice sheet, since the interior seems chiefly to be nourished by snowfall from this anticyclone, and since, furthermore, low temperature near the ground is the essential condition for precipitation in a glacial anticyclone (Simpson, 1919, pp. 256–269), a lowering of temperature may have been the cause of the heavier precipitation. Moist sea breezes could

hardly have reached the central parts of the ice sheet with any regularity even during the postglacial temperature maximum. Considering the latter alternative to be the more probable, the greatest extent of the ice may be correlated with the heaviest glaciation in South America and other parts of the southern hemisphere. Furthermore, the sheet may have been less extensive than now during the postglacial maximum of temperature.

### Northern and Southern Hemispheres

The Quaternary drift deposits in South America and in Australia, according to the unanimous testimony of those who have studied them, correspond in general appearance, weathering, and erosion with the glacial formations in the northern hemisphere (Nordenskjöld, 1898, p. 71; Meyer, 1904, p. 598; 1907; Steinmann, 1906, p. 228; Hauthal, 1911, p. 182; Coleman, 1924, p. 400; Klute, 1925; Gregory, 1904, p. 52; Süssmilch, 1911, p. 152; David, 1923, p. 130; Clemes, 1924, p. 69). Pleistocene glaciation heavier than the present extended straight across the equatorial region of South America. Indications of lower temperature during the Pleistocene have been found in all regions of the earth that have been studied. Corresponding late-Quaternary climatic changes seem to have taken place in both polar hemispheres. All these conditions make perfect or approximate contemporaneity of the glaciations on the northern and the southern hemispheres highly probable, yet they hardly prove it. Observations on the drift and on the erosion of its surface are too few and too subjective, and alteration processes are too little known.

The climatic relation between the hemispheres is not well understood. It is doubtful whether such great climatic changes as those that led to glaciation could, with unchanged position of the poles, occur in one polar hemisphere without also profoundly affecting the other. The chronology is only approximate. At least there could well have been a slight difference in time between the greatest development of the glaciers under high latitudes on opposite sides of the equator. The glaciers

under and near the equator could have had their maximum extent simultaneously with the climaxes of the glaciations on both hemispheres.

While the actual time relation between the climaxes on the northern and the southern hemisphere cannot be determined by comparing the depth of leaching or the physiographic changes, nor by climatologic speculations, it perhaps can be determined by studies of the retreat of the ice sheets and glaciers from their last great extent. If the waning of the ice in the southern hemisphere presents exactly the same *long-range* periodicity as it does in the northern, full contemporaneity may be concluded. Dr. Carl Caldenius' geochronological studies in Argentina and Patagonia will perhaps ultimately solve the problem (cf. De Geer, 1927).

# CHAPTER II

## CLIMATIC CONDITIONS DURING GLACIATION

In considering climatic conditions we may divide the glacial period into three stages: (a) that of growth and extension of the ice sheets; (b) that of the maximum stand and construction of the outermost moraine; and (c) that of decline and disappearance. Of these stages, the first is obviously the most difficult to understand, because of its strange requirements and the scanty evidence upon which to build. The second, to be regarded probably as a transition between the other two, is thought to be of comparatively short duration. The third or closing stage is more fully revealed by a great variety of records.

Direct and indirect indications of the conditions which led to the waxing and waning of the ice sheets are found in snow lines of the past and present, in temperature and rainfall requirements of the known ice masses at different altitudes, in the flora and fauna found as fossils in the drift deposits, in the surviving colonies of polar species both in and outside of the glaciated areas, and in the record of varved clays that register varying rates of recession and the duration of temporary halts and readvances during final retreat. All these will be considered at appropriate points in the chapter.

### Climate during Stage of Growth

#### CHIEF FACTORS OF GLACIATION

The two chief meteorological factors of glaciation are low summer temperature and heavy precipitation in solid form. The importance of low summer temperature is evident from the universal indications of it during the Pleistocene glaciations, from the important rôle it played in late-glacial and postglacial climatic changes, and from various other conditions. In the

Alps the gathering fields during the Pleistocene contained no more snow or névé than at the present, showing that the expansion of the glaciers at that time was not due to heavier precipitation but to diminished ablation of the lower parts of the glaciers because of decrease of temperatures above melting (Penck and Brückner, 1909, pp. 1142, 1145; Brückner 1910, p. 106; 1912, p. 389). Also the practically complete parallelism between the Pleistocene and the present snow lines shows that the cause of the Pleistocene depression of the snow line must have been low summer temperature, while the precipitation was about as at present; for it is inconceivable that an increase in precipitation could be such that the snow line everywhere in the large area would fall the same amount (Brückner, 1910, p. 107). Furthermore, if the cause of the lowering of the snow line be supposed to have been augmented precipitation, this increase must have been unreasonably great. The necessary snowfall is estimated at 20 to 30 m. (Brückner, 1914, p. 195).

The altitude of the snow line is determined first of all by temperature, this being subordinate only in regions with very abundant snowfall (Paschinger, 1912, p. 85). According to Köppen's (1920) revision, temperature is decisive for the snow line in mid-latitudes while precipitation means most for it in low latitudes. That the summer temperature was the primary factor for the growth of the ice sheets may also be concluded from the fact that it was the chief cause of their waning (Antevs, 1925b, p. 48). That it is actually absence of summer heat, a minimum of temperatures above thaw, and not low winter temperature nor low annual temperature, that produces glaciation is also distinctly shown by a compilation by Köppen (1924, p. 200) on the relation between temperature and glaciation of some regions below the 65th parallel. Eastern Siberia, in spite of its extremely cold winters, is not glaciated; while southern Greenland, lying on the same latitude, is covered by an ice sheet. The annual mean temperature (at the same altitude) is 8° C. (14.4° F.) lower in Siberia, but the average summer temperature here is 7° C. (12.6° F.) higher, being 12° to 20° C. (53.6° to 68°

F.). The precipitation is small in both regions. It should be remembered, though, that the Greenlandic ice is a survival from the Ice Age, in which supply and wastage just balance each other.

However, summer temperature is not unanimously considered to play the chief rôle in the climatic changes leading to glaciation and deglaciation. Precipitation is thought by Brockmann-Jerosch, von Drygalski, Huntington, and Visher to be of greater importance.

Brockmann-Jerosch (1910, 1919, 1920) attributes the Pleistocene glaciations to abundant snowfall and even thinks that the temperatures during the Ice Age were not essentially different from those now prevailing. He emphasizes the fact that the glacial Dryas flora is mixed with aquatic plants and swamp plants which require a relatively long vegetative period and now extend neither up in the Alpine zone nor poleward on the farther side of the timber line. However, Brockmann-Jerosch does not distinguish between the stage of expansion and stage of waning of the ice. Most finds of Dryas flora date from the stage of ice retreat which normally was characterized by fairly high summer temperature. During this age shallow ponds in the vicinity of the ice border could be considerably heated by insolation during the summer months, as is also shown by remains of fresh-water molluscs; and thus temperate aquatic plants could thrive here while arctic plants lived on the surrounding land. From the mixed late-glacial flora evidently no conclusions can be drawn regarding the climatic conditions during the growth of the ice sheets.

von Drygalski (1919, pp. 31–35) starts from the unquestionable fact that the amount of snowfall is decisive for changes in the extent of the Antarctic ice sheet. He thinks that during the Pleistocene there was an increase in precipitation all over the globe. He discusses particularly the Pleistocene glaciations in the Alps. Against Penck's and Bruckner's tendency to attribute the former extensive glaciations in these mountains to fall of temperature it is argued that the névé fields could have contained more snow and firn than they do today without leaving traces on the steep mountain walls and that a snowfall

corresponding to 11 to 14 meters of water, calculated to be necessary for the Pleistocene expansion of the glaciers with temperatures as now, is a maximum and not an average figure.

However, the Antarctic is an unfortunately chosen area to start from, since in this continent low summer temperature, the very factor whose importance von Drygalski doubts, prevails more conspicuously than anywhere else. The isotherm of 0° C. (32° F.) for the warmest month of the year roughly coincides with the polar circle (Ward, 1918, p. 162). Furthermore, probably all precipitation comes in solid form, and thaw occurs only locally and at very infrequent intervals (Wright and Priestley, 1922, p. 280). The objections against Penck's and Brückner's view seem to carry little weight. There is no denying that unreasonable quantities of snow must be assumed to explain the Pleistocene expansion of the Alpine glaciers and still greater quantities to explain the growth of the North American ice sheet without a primary fall of the temperature. The assumption of a synchronous world-wide increase in precipitation seems doubtful, for, if stronger winds would tend to hasten evaporation, the temperature fall, which must have been caused by the growing ice sheets, would tend to lessen it.

Huntington and Visher (1922; Huntington, 1925) also hold that precipitation—or, better, storminess—plays a greater rôle than temperature. In his latest treatise on the subject Huntington shows particularly that a lowering of the earth's mean temperature does not tend to increase the snowy precipitation but rather to diminish it. The effect of temperature upon snowfall is drawn into the discussion because ice sheets are supposed by Huntington to be nourished by snowfall from cyclonic storms. This kind of snowfall no doubt played a rôle but probably only a subordinate one. Over the vast ice sheets there seem to have existed strong anticyclones, so that storms could encroach only a little on the fringe of the ice. In the Antarctic snowfall from moist winds blowing in from the sea is limited to the border belt (Wright and Priestley, 1922, p. 24). On the other hand the chief nourishment of the Pleistocene ice

shields may have taken place in the interior parts, from where the ice sheets slowly grew out. This is probable from the outward transportation of part of the fallen snow, the outward slope of the ice surface, and the centrifugal ice motion recorded by long carriage of boulders (see p. 65). The excentric growth of the ice sheets, the incomparably greater extension southward and landward than in other directions may have been due to development of low pressures over the continents in summer time, strengthening the anticyclonic winds towards them, and to easier flow of the ice here because of higher summer temperature. Without heavy snowfall and partial consolidation to ice in the interior the ice sheets would have been thin there and eventually have consisted of a ring around an ice-free center. They would have lacked motion.

The mode of snowfall in the glacial anticyclones may perhaps be supposed to have resembled that in the modern ones as interpreted with but slight difference of view by Hobbs (1911, 1921, 1926) and Simpson (1919, pp. 256–269; also Wright and Priestley, 1922, p. 9). In the Antarctic the outblowing surface air is very cold in the beginning because of radiation, contact with the ice, and other factors. Pressure waves pushing from behind force it to rise to moderate height below the inflowing upper air current and it is rapidly cooled further in the ascent, so that the water contained is precipitated as snow. Thus the low temperature of the lower strata of the air in the interior of the ice sheets may have promoted snowfall on them.

However, various writers hold that nourishment of the ice sheets under permanent anticyclonic pressure cannot take place. van Everdingen (1926) and Exner (1926) think that the solution of the problem lies in changes in the position of the anticyclones. "At times when the whole anticyclone is removed, fairly warm air from sea level can cover the entire land area and already during rise on slopes cause some precipitation. If the anticyclonic pressure is restored under influence of air currents at high levels, the cyclonic air will be cooled off and give up its moisture" (van Everdingen; also Hörner, 1927, p. 180).

Cyclonic storms are assumed by Huntington to account not only for the nourishment of the ice sheets but also for the lower temperature, but it is not easy to understand how they themselves could cause sufficient fall of the summer temperature, so that the Pleistocene expansion of the glaciers in regions with considerable present snowfall, such as the Alps and northern Europe, can be explained without resort to unreasonably large quantities of snowfall.

Precipitation, however, was by no means unimportant. In fact, locally, in cold, unglaciated regions, like Labrador and Keewatin, increased snowfall must have been essential for glaciation. But glaciers originating there under temperature conditions like the present could not possibly extend down to New York City, as secondary cooling of the summers would be entirely insufficient to prevent complete melting under the lower latitudes. A primary fall of the summer temperature there is necessary.

Thus, since in some regions with abundant precipitation drop in summer temperature seems to have been the sole factor of glaciation and since even in the glaciated areas whose centers have deficient precipitation primary fall of temperature is a necessary condition for the large expansion of the ice sheet, low summer temperature may be regarded as the most prominent factor in glaciation, with heavy snowfall playing a very important rôle. The growth of the ice sheets in low latitudes evidently demanded considerable nourishment.

## CLIMATIC CONDITIONS

The stage of growth of the Pleistocene ice sheets was characterized by excess of nourishment over wastage. This effect could be attained by various relations of the two chief controlling factors, viz., summer temperature and snowfall. In regions that already had heavy precipitation only a slight temperature fall was needed. In other areas both increased snowfall and decreased summer temperature were required. At high latitudes with small summer depletion the need of nourishment was but slight. In low latitudes with long summers and strong insolation it must have been great.

Thus, primarily the summer temperature was lower than at present in the same region. Since the temperature was lower by almost equal amounts at the different latitudes, the precipitation must have been increasingly greater with the lower latitude to counterbalance the greater wastage. The primary temperature fall probably had only to be slight. The amount of lowering of the summer temperature during the climax of the last glaciation has, for instance, been estimated at 4° C. (7.2° F.) in the Alps (Brückner, 1910, p. 108) and 5° C. (9° F.) in Australia (Speight, 1921a, p. 346). At points well inside the ice edge the temperature lowering evidently was very much greater.

Through primary lowering of the summer temperature or increase of the temperatures below thaw a greater part of the precipitation came in solid form, the amount of snow melting decreased, and the snow line in the mountains moved downwards. In many cases increase in snowfall was necessary and occasionally was more important than temperature fall. Radiation from the permanent snow and ice cover cooled off the overlying air, which tended to flow out towards the snow-free land. There also was a temperature decrease of the air in the surrounding area. As the ice field grew in extent there developed a local anticyclone sufficient to modify the general air circulation in its neighborhood. This high increased in strength with the growth of the ice sheet, and the nourishment largely took place through snowfall in the glacial anticyclone and to a lesser extent by snowfall from moist winds blowing in over the peripheral parts. The strong highs over the ice were perhaps permanent and postulate permanent lows over the northern Atlantic and the northern Pacific. Cold winds sweeping down from the ice fields were a typical feature. The climate became most severe during the last stage of waxing of the ice.

The climate was rather different from that now prevailing under high latitudes and more comparable to that in high mountains under mid-latitudes. The chief differences were connected with the latitude, for there does not seem to be sufficient reason to assume that the poles during the Pleistocene

glaciations held other positions than today. Under the low latitudes the summers were long, if not so long as now; the angle of incidence of the sun's rays was great; and the temperature in the sunshine may consequently have been fairly high. Inside the polar circles the summer is short, but the sun does not set in summer for a period of from one day on the polar circles to six months at the pole, and the summer days are on the whole very long. The sun's rays are very oblique and increasingly so poleward, and the atmospheric absorption, dependent upon the obliquity of the sun's rays, is great and the temperature in the sunshine actually low, even if the direct insolation is very effective.

As the glaciers expanded, animals and plants inhabiting the affected areas were forced to migrate down the mountains and towards ice-free land. The north-European flora and fauna retired to central and southern Europe, and the flora and fauna of the Alps and other mountains withdrew to the surrounding lowlands. So the lowlands of Europe became the refuge and scene of intermingling of different floras and faunas (Nehring, 1890; Zschokke, 1902, 1912; Ekman, 1922, p. 533; Braun-Blanquet, 1923; Lämmermayr, 1923). However, not all biota undertook migrations. A few plants survived the glaciation even in the central ranges of the Alps (Braun-Blanquet, 1923, p. 253). In the ice-free regions on the Norwegian coast outside the continuous ice sheet and on nunataks rising above the ice in Norway also a number of plants and animals weathered the last glaciation (see p. 26). During the repeated glaciations in Quaternary time a number of Pliocene deciduous trees that could not escape across the Alps, the Pyrenees, and the other east-westerly mountain ranges to regions with more genial climates were exterminated. For this reason the modern European forest is poor in species of foliiferous trees as compared with that of North America where there was no barrier for the biota that were driven from their old homes.

The growth of the ice in Keewatin and Labrador must have required much heavier snowfall than the present. Consequently

there must have been considerable transportation of moisture from the sea in low and middle latitudes. The unglaciated regions probably received less rain than they do at the present time, because with lower temperature evaporation probably was not greater than it is now.

## Climate at the Time of Maximum Glaciation

### TEMPERATE BELTS

### Central Europe

*Tundra Belt*—The ice sheets expanded, finally reached their limits, and ceased to grow. As the subsequent shrinking surely was due to a rise in summer temperature and a decrease in snowfall, it is probable that this same climatic change was the cause of the cessation of growth. At any rate, wastage counterbalanced supply. The climate just before, at, and just after the attainment of the greatest extent is indicated by the subfossil flora and fauna and less directly by other evidence.

The floras and faunas were largely arranged in zones. Just off the edge of the ice, as the remains embedded in the glacial drift and at the bottom of the peat bogs show, there was a vegetation consisting largely of mosses and lichens but also of *Dryas octopetala*, *Salix polaris*, *S. herbacea*, *S. myrtilloides*, *Betula nana*, etc. Forest trees were lacking. True arctic floral elements spread as far south as the Pyrenees (in one instance to central Spain), the Abruzzi in central Italy, and the northern part of the Balkan Peninsula as shown by the modern occurrence of relict forms. They did not reach the Caucasus (Kulczyński, 1924, pp. 162, 165, 167). The birch and the pine, which at present extend to the arctic timber line, may have occurred during the climax of the glaciation in Bohemia (Rudolph and Firbas, 1924), though they are thought by H. A. Weber (1919, p. 243) to have been lacking in southern and central Germany.

In this belt with high-arctic vegetation there lived the banded lemming, the Obi lemming, the Arctic fox, the reindeer, the musk ox, etc. All these animals are characteristic of the arctic tun-

dras or barren grounds, especially the two lemmings, which are confined to this type of land (Nehring, 1890, p. 21; Haviland, 1926). Remains of this fauna are found all over central Europe including the Alps as far west as France, northernmost Spain, and England (Geikie, 1914, p. 23; Harlé, 1910; Osborn, 1915, p. 303). Hence during the climax of the glaciations the greater part of the region between the Scandinavian ice sheet and the Alps, as has already been so convincingly shown by Nehring, was occupied by tundras in most respects comparable to the present tundras of the high latitudes.

The climate of the modern tundras is characterized either by cool summers or by short, mild summers and very cold winters with scanty snowfall (Passarge, 1921, p. 56). That of the Eurasian tundras is on the whole characterized less by cold winters, though these are very cold, than by cool summers (Kendrew, 1922, p. 187). The ground is frozen during most of the year and at a few feet below the surface is frozen permanently. In summer the surface thaws for a short time but is then water-logged except on south-facing slopes. The precipitation is small, but the air is damp and raw. During the coldest time the air is remarkably still, but especially toward the end of winter violent drying-out wind storms frequently sweep the flat, unprotected spaces. The southern limit of the north European tundra coincides nearly with the July isotherm of 12° C. (53.6° F.) or perhaps more closely with a line combining points having a summer period 1½ months long during which the mean temperature is 10° C. (50° F.) and above (see Werth, 1925, p. 396).

The climate of the glacial tundras of central Europe may have been essentially the same as that just described. It was continental with slight precipitation (C. A. Weber, 1914, p. 63). However, the lower latitude and the ice sheet made some differences. The summer was longer, the temperature of both summer and winter was probably higher, and cold winds may have swept down with fair regularity from the ice sheet. Mechanical weathering testifies to low winter temperature; solifluction, to low temperature, water-logged ground, and a sparse cover of

vegetation (cf., however, W. Penck, 1924, p. 88); and deposition of wind-blown loess to aridity. Relatively warm summers, very cold winters, large daily and unperiodic waves of temperature, icy cold drying winds from the ice sheet, scanty precipitation, and great evaporation are by Kessler (1925, p. 179; 1925a) held to be especially characteristic. In Thüringen and Sachsen the mean temperature during two months in summer may have attained 10° C. (50° F.) or over (Werth, 1925, p. 397). The severe winters may have necessitated annual migrations of part of the fauna.

During the climax the Dryas zone was quite wide, but as the ice waned it decreased in width and finally became discontinuous as a distinct belt in central Sweden (H. A. Weber, 1919, p. 239; von Post, 1911, p. 20).

*Steppe Belt*—Remains of a fauna very different from that of the tundra show that that belt did not occupy all of unglaciated western and eastern Europe, even if it did occupy the whole area directly between the ice sheet and the Alps (Nehring, 1890, pp. 172–235; Stoller, 1910, pp. 173, 181; Krause, 1910, p. 124; Geikie, 1914, pp. 25–32). As the ice began to retreat this fauna spread northward. It consists of jerboa (*Alactaga jaculus*), red suslik (*Spermophilus rufescens*), bobac or steppe marmot (*Arctomys bobac*), saiga antelope (*Saiga tatarica*), and several other forms characteristic of the modern steppes of eastern Russia and western Siberia. The fossil remains occur much farther west than the living fauna, partly even in France, but not in Spain or Portugal (Osborn, 1915, p. 303) nor in the Scandinavian countries. *Trisetum Cavanillesii*, *Poa concinna*, several *Astragalus* species, and other plants at present mainly found on the steppes on the Danube and in southern Russia, were dispersed westward and live today as refugees in warm and dry valleys of the central Alps (Braun-Blanquet, 1923, p. 225). From these conditions it is inferred that in central Europe the tundra belt was bordered by grass steppes and park steppes.

The climate of the modern steppes and most probably also of the late-glacial steppes is continental (Passarge, 1922, p. 55).

In summer the days are warm, the nights cool. The winters are cold, often very cold, with disastrous snowstorms. Change from winter to summer is surprisingly rapid. Precipitation is unevenly distributed over the year and, especially in summer, is insufficient, so that vegetation is parched and burnt up, leaving the sand and dust as prey to the sweeping warm winds. Occasionally the precipitation is excessive.

*Forest Belt*—Outside the steppe belt there lay a forest zone, as recorded by buried vegetation in the peat bogs. That the three belts were not sharply separated one from another, is shown by finds of mixed floras and faunas. Both plants and animals of the distal belts invaded the more proximal ones before the disappearance of the characteristic biota. Trees formed a typical feature of the park steppes and probably were not entirely lacking even on the tundras. The cold region between the Scandinavian and the Alpine ice caps divided the forest and its fauna into a small west-European part and a larger eastern part (H. A. Weber, 1919, pp. 239–243; Braun-Blanquet, 1923, p. 251). The forest trees, of course, had a zonal distribution. The first trees of importance taking possession of the previous steppe or tundra were birch (*Betula odorata* and *B. verrucosa*) and aspen (*Populus tremula*), while pine (*Pinus silvestris*) followed soon after. Forests not too far from the extreme border of the ice, as in northeastern France, Poland, and Galicia, consisted of *Pinus silvestris*, *P. Cembra*, and *Larix decidua* with northern and Alpine herbs (Gams, 1927).

## Western Norway

The distribution of some arctic plants and animals in Scandinavia makes it highly probable that they survived the last glaciation here and that islands and mountain peaks on the greater part of the western coast of Norway and also parts of the Dovre and the Jotunheimen Mountains escaped being buried under the ice (Blytt, 1893; Wille, 1905, 1915; Ahlmann, 1919, p. 217; Ekman, 1922, pp. 397–404). The chief districts of refuge were the outer parts of the Lofoten and Vesterålen Islands.

The surviving flora consisted of willow (*Salix polaris*) and similar plants.

Among surviving animals is especially to be noted the Norwegian lemming (*Myodes* or *Lemmus lemmus*) (Ekman, 1922). This lemming did not live in central Europe during or after the last glaciation, and the species cannot very well have arisen in postglacial time from the Obi lemming (*Lemmus obesis*), since both these lemmings are distinct species and in the Scandinavian vertebrate fauna there is no other comparable example of formation of species during the postglacial epoch. The modern distribution of the Norwegian lemming comprises the Scandinavian mountain chain and the Kola Peninsula, and the present western limit of the Obi lemming lies to the east of the White Sea. Besides the Norwegian lemming probably various other kinds of animal life, such as arctic seafowl, butterflies, etc., existed in Norway during the glaciation, though no conclusive evidence of this can be given.

The survival of plants and animals in unglaciated tracts and on nunataks within the glaciated area, of course, is quite to be expected. In a small nunatak in northernmost Greenland (81° N.) 8 phanerogams are recorded; and on Jensen's Nunatakker (62¼° N.) in western Greenland, 70 km. (44 miles) inside the ice border and at an altitude of 1700 m. (5600 feet), 27 phanerogams have been found (Ostenfeld, 1926, p. 12). As is commonly known, almost all arctic and antarctic lands have their floras and faunas, even though in many cases they are very poor as compared with those of lower latitudes. A floristic province comprising part of northernmost Greenland and forming an isolated strip of land between the 80th and 84th parallels contains 62 species of higher plants (Porsild, 1921, p. 47). In all Greenland 30 mammals are known (Stephensen, 1921, p. 58). Certain regions have a surprising abundance of biota. In Franz Josef Land, which is almost entirely ice-covered and is the most decidedly arctic land on the northern hemisphere, 23 vascular plants are known, while 125 are known in Spitsbergen (Supan, 1921, p. 835; Nordenskjöld, 1909, p. 59). In what is probably

the most extreme case, the Antarctic, there is almost no land vegetation, only a few species of mosses and lichens being known; while in the most favored parts outside the antarctic circle, there are one grass (*Descampsia antarctica*) and one dicotyl (*Colobanthus crassifolius*). No real land animals of the higher order are known (Nordenskjöld, 1909, p. 94; Brown, 1906, 1912; Skottsberg, 1921). The absolute scarcity of land biota may, however, to a great extent be due not only to the cold Antarctic summer but to extermination during the climax of glaciation here and to inability to reimmigrate later across the wide surrounding seas.

Western Norway during the last glaciation may not have been more inhospitable than northern Greenland today. Biota may have survived under much higher latitudes, as in Spitsbergen. However, the reindeer seems to cross the ice from Nova Zemlya, for several marked animals have been killed in Spitsbergen (Wollebæk, 1926, p. 60). In Greenland some 60 species of vascular plants may have survived (Ostenfeld, 1926, p. 13).

## Northern Asia

The extent of the last mountain glaciation in Asia (see p. 4), apparently synchronous with the ice sheets in Europe and North America, shows that the summer temperature was then essentially lower than at present. This is attested also by the floral exchange that then took place between the arctic tundra and the high regions of the Altai (Kulczyński, 1924, p. 166) and the extensive occurrence of alpine and sub-alpine plants in widely separated mountain ranges in China (Limpricht in Olbricht, 1923).

The precipitation is mostly believed to have been about the same as today. This seems to be indicated, for instance, by the parallelism between the Pleistocene and the modern snow line in the Altai and the relatively limited glaciation. However, the Pleistocene lowering of the snow line here is explained by Fickeler (1925, p. 128; 1925a, p. 633) as a consequence of greater precipitation from winds coming from the same direction as those which bring moisture at the present time. (Cf. p. 39.)

## Eastern North America

The conclusions about the climate that can be drawn directly from the existence of the ice sheet have been briefly touched upon on page 20. The fossil remains, the modern distribution of the biota, etc., indicate that conditions much like those in Europe prevailed in North America. The ice sheet apparently was bordered by a belt with plants and animals characteristic of the modern tundras or barren grounds of the north. Of the musk ox, now entirely confined to arctic North America and Greenland, fossil remains probably dating from the last glaciation have been found outside the outermost moraines in Indiana, Illinois, West Virginia, Missouri, and Iowa and within the glaciated area in Ohio and Indiana (Hay, 1923, pp. 248–255; 1924, pp. 178–185). Of the reindeer, remains also probably originating during the last glaciation have been found outside the terminal moraines in New Jersey and at various places inside them as at New Haven, in Vermont, and at Toronto (Hay, 1923, pp. 244–247). However, though essentially an inhabitant of the tundra, the reindeer in modern time occurred down to the latitude of the Great Lakes (Seton, 1910, p. 192) and lived in Germany in historic time (Ekman, 1922).

The presence of such northern forms as the white pine (*Pinus strobus*), spruce (*Picea mariana*) and hemlock (*Tsuga canadensis*) in the southern Appalachians may be a consequence of the glacial period (Gray, 1878). The occurrence in Nebraska of branches, twigs, and occasionally trunks of conifers at depths of 25 to 30 feet indicates the existence of extensive forests in not remote geologic time (C. E. Bessey, see Transeau, 1903, p. 413). The present climatic conditions are too dry, and the coniferous forest has withdrawn to the northeast of the state. The bog plants sundew (*Drosera*), pitcher plant (*Sarracenia purpurea*), tamarack (*Larix laricina*), cranberry (*Oxycoccus oxycoccus*), cassandra (*Chamaedaphne calyculata*), Labrador tea (*Ledum groenlandicum*), etc. in Ohio, Indiana, southern Michigan, etc. are late-glacial relicts, their modern real area of occurrence being a belt north of the Great Lakes (Transeau, 1903).

As the ice border retired, parts of the arctic flora and fauna withdrew up the highest mountains. Isolated from their real areas of distribution, they have here been able to survive till the present day, testifying to conditions outside the terminal moraines during the maximum extent of the ice sheet. Arctic relict biota are especially to be found in the White Mountains of New Hampshire, the Adirondacks in New York, on Mt. Katahdin in Maine, and in the Green Mountains of Vermont. On the upper slopes and summits of the White Mountains (6293 feet; 1918 m. high) there occur 55 vascular plants of Greenland and Lapland, among which may be mentioned *Phleum alpinum*, *Salix* spp., *Rhododendron lapponicum*, and *Diapensia lapponica* (Fernald, 1925, pp. 266–269, 293). In addition there are to be found a number of high arctic invertebrate animals, as a black spider—otherwise confined to Labrador, Greenland, and the Rocky Mountains—a butterfly, moths, beetles, etc. (see Scharff, 1912, pp. 36–37). On Mt. Marcy (5344 feet; 1629 m.), the highest peak in the Adirondacks, where an alpine zone with stunted trees is to be found from the elevation of 4900 feet (1495 m.) to the summit, a flora and a fauna with a smaller percentage of arctic species are met with (Adams and others, 1920, pp. 216–230). The flora, which probably contains 23 true alpine forms, includes *Diapensia lapponica* and *Rhododendron lapponicum*. The invertebrate fauna is poor. Its biology and environmental conditions are too little known to permit the drawing of definite conclusions regarding the late-glacial climate (Adams and others, 1920, p. 232).

While arctic biota largely migrated in front of the growing ice sheet to central North America and then retreated northward as the ice melted away, the same flora and fauna to some extent survived at least the last glaciation in the region of the Gulf of St. Lawrence, in the Torngat Mountains of northeastern Labrador, and in the American Arctic Archipelago (Fernald, 1925). The districts of survival on the St. Lawrence Gulf are the Shickshock Mountains of the Gaspé Peninsula, the Long Range of western Newfoundland, and the Magdalen Islands. These latter areas and the Torngat Mountains were undoubtedly ice-

free during the last glaciation (Coleman, 1920). The extent of eventually ice-free land in the Arctic Archipelago is unknown (see p. 76), but probably lesser areas escaped glaciation than is believed by Fernald (1925, p. 325). Since their uncovering from the ice the greater part of northern Baffin Land, i.e. up to 1300 feet (400 m.) of altitude, the whole of Melville Peninsula, and perhaps all of Southampton Island have been submerged beneath the sea (Freuchen and Mathiassen, 1925, pp. 554, 555, 560). There may not have been such an eastward migration of the glaciated areas with only a relatively small part of the width of the continent glaciated at each time as is assumed after Tyrrell (Fernald, 1925, pp. 245, 326; cf. our p. 2). In spite of this the biota may have been able to survive the glaciation (cf. p. 27).

The tundra probably formed only a narrow zone on the margin of the Pleistocene ice sheet. While in central Europe the tundra was bordered by steppes, this does not seem to have been the case in North America, though there are loess deposits indicative of a somewhat drier climate than the present (Alden, 1910; Gleason, 1922). Outside the tundra belt there "came a third belt of still less homogeneity; in the east it was composed of deciduous forests and their associated fauna, while in the west it was made up of plains and desert types of life" (Adams, 1905, p. 56).

## Alaska

The Pleistocene climate of Alaska like that of Siberia is commonly thought to have been extraordinary. In the Quaternary beds, for instance in those of the Kuzitrin lowland (65° N., 165° W.) on Seward Peninsula, in regions which now are treeless there have been found together with remains of the mammoth, the horse, and other mammals large trunks of trees (Collier, 1908, pp. 89, 91; also Maddren, 1905; Quackenbush, 1909, p. 126; Scharff, 1912, pp. 78–80). On the other hand the fossil land biota, as far as they are known, do not furnish any direct evidence for a colder climate during any part of the Quaternary than that now prevailing. From these facts the conclusion has been drawn

that, even when great parts of the North American continent were buried beneath ice sheets, Alaska enjoyed a milder climate than it does today and that in postglacial time a deterioration set in causing extermination of part of the fauna and forcing the forest southward. This explanation seems improbable. The Quaternary, besides cold periods, also had warm interglacial epochs; and the warm-climate beds no doubt date from these. The mammoth and the horse probably died out during one of the glaciations, when the temperature surely was considerably lower than at present, for the extent of the Pleistocene ice caps and glaciers in Alaska was by no means insignificant (Brooks, 1906, p. 295, Pl. 22). That the temperature was lower during the last one is directly shown by the then marine molluscan fauna on the coast from Oregon northward (Dall, 1907).

Since the climate of Alaska during the interglacial epochs may have been essentially arctic and subarctic, as it is now, its flora and fauna could not very well undergo great changes from interglacial to glacial age. It seems probable, though, that systematic stratigraphic studies of the deposits will reveal more evidence than is here quoted, of retreats and advances of the biota as the ice caps waxed and waned.

## THE PRESENT ARID BELTS

The drop in temperature and the gradually increasing anticyclones above the ice sheets greatly changed meteorological conditions also in far distant regions. The climax of these changes was naturally reached at about the time of the maximum extent of the ice. In the unglaciated regions the climax may have occurred simultaneously with the great climatic change that put an end to the growth of the ice sheets. This is probable because abnormally great precipitation could not well occur at the same time over the ice sheets and in other regions of the globe on account of the low temperature and consequently not exceptional evaporation. The decrease in precipitation that now began over the ice sheets was correlated with an increase in rainfall in the belts extending equatorward from the ice caps. Because of this and

because of less evaporation due to lower temperature the greater part, the polar part, of the modern arid and semi-arid belts extending in the northern hemisphere over northern Africa, the Mediterranean, Asia Minor, Syria, Arabia, Persia, Turkestan, Mongolia, and the Basin-and-Range Province, and in the southern hemisphere over southwestern Africa, most of Australia, and a large area in South America (see Köppen, 1923; Passarge, 1924) enjoyed moister conditions than they do today. This is shown by evidence of various kinds, among which the former vast extension of bodies of water, the great watercourses now dry, and fossil occurrence and present distribution of plants and animals are the most important.

The arid conditions that now prevail in the regions enumerated are due to great evaporation; to large expanse of land that makes it difficult for moist winds to reach the interior; to high mountain ranges shutting out air currents that bring rain and cause moisture to become precipitated; to cold ocean currents that reduce the capacity of the air for holding vapor so that this increases when the air moves over heated land and evaporation instead of rainfall results; and to the atmospheric pressure distribution with its consequent prevailing winds. The aridity in the belts between about 20 and 32 degrees of latitude nearly all round the globe except in the monsoon regions of Asia, is chiefly due to the planetary arrangement of pressure, to the subtropical high pressures at 20 to 40 degrees of latitude, and to the trade winds which blow from these towards the equatorial low-pressure belt. Anticyclones are systems of descending air currents: as the air sinks it becomes compressed and thus warmed and dried; its vapor-holding capacity increases; the sky is cloudless; no rain falls; but, instead, the winds are dry. Thus, the high-pressure belts are arid. The trade winds, also, are dry winds both because of their origin and because of the fact that they reach warmer and warmer regions so that their capacity for vapor steadily increases. As the pressure belts shift northward and southward with the sun, the arid zones are bordered on the north and the south by semi-arid belts with seasonal rainfall.

The climatic conditions in these belts during the glacial-pluvial epoch, as A. Penck (1914) first pointed out, were due to heavier rainfall, to smaller evaporation because of lower temperature and to displacement of the pressure zones towards the equator.

The contemporaneous climates on the equatorial border of the deserts seem to have been locally moister or drier than today. The Mexico Basin was moister; the southern border of the arid belt lay here farther from the equator than now, so that the desert zone was narrowed both from north and south (Jaeger, 1926, p. 62). On the other hand the desert in southwestern Africa and northern Australia seems then to have extended nearer the equator than today (pp. 43, 45).

As the temperature rose and the ice sheets waned the climatic belts of low and middle latitudes expanded and migrated poleward to about their modern positions. The arid belts, in Berg's (1913) opinion, reached their greatest extent in postglacial time and in modern times are shrinking both on the equatorial side and the polar side.

## The Mediterranean and Northern Africa

The aridity of the Sahara is essentially due to the persistence of dry northeasterly trade winds throughout the year. In summer the arid conditions move northward over the Mediterranean, in winter southward over the Sudan, almost to the Gulf of Guinea (Kendrew, 1922, p. 24). The mountains in the desert receive much more rain than the plains. The Ahaggar receives both winter rains from the north and summer rains from the south, while the southern ranges, the Air and the Tibesti, are reached only by rain-bringing winds from the southwest.

Mediterranean Africa has very warm and practically rainless summers with prevailing northerly winds. In winter it has westerly winds and abundant rainfall, associated with cyclones passing from west to east off the coast. The northern part of the Mediterranean has northerly winds throughout the year owing to continuous lower pressure over it than on the European continent. The summers are hot and dry, the winters mild. Autumn and spring bring most rains.

Conditions during the pluvial epoch, which will presently be discussed, prove that the subtropical belts of high pressure and the temperate belts of low pressure lay much farther south than today and that rain-bringing cyclones traveled eastward across the Mediterranean throughout the year.

Pleistocene glaciation occurred all around the Mediterranean on mountains rising about 2500 m. (8200 feet), as on the Great Atlas of Morocco (Gentil, 1924; de Martonne, 1924), on various ranges in the Iberian Peninsula (Obermaier, 1921), in Corsica (Lucerna, 1910), in Italy (Taramelli, 1910; De Lorenzo and Dainelli, 1923), in the Balkan Peninsula (Cvijić, 1908; A. Penck, 1925b, pp. 851, 854, 855, 858–860), and on Mt. Olympus in Asia Minor (Cvijić, 1908).

Fossil remains of the moisture-requiring *Rhododendron ponticum* in the island of Skyros, Greece, are conclusive proof of greater humidity at the time of its existence (G. Andersson, 1910, p. 149). The relict occurrence of this plant at a few localities in the southern part of the Iberian Peninsula suggests a formerly moister climate here also. The greater humidity was perhaps due rather to small evaporation on account of low temperature than to heavy rainfall, as the distribution of the Pleistocene glaciers suggests a precipitation of about the same amount as at present (Obermaier, 1921, p. 162).

In the Sahara, particularly in the French part between the Atlas Mountains and the Niger River, there are extensive wadis or river channels (Chudeau, 1921, 1921a; Gautier, 1923, pp. 48–54). From the Ahaggar massif in central Sahara old channels diverge in all directions, and from the Atlas a net of fossil rivers trend toward the south and southeast. The occurrence of tropical fishes in water holes and subterranean waters of the Biskra oasis and of Wadi R'ir, the presence of Ethiopian fresh-water molluscs and of the crocodile in the same regions, of the Indian cobra in southern Algeria, etc. bear witness to a time when there was a communication of running water between the equatorial and the Mediterranean regions. Communication was established by the Pleistocene rivers Tafassasset, flowing from the Ahaggar

to the Gulf of Guinea, and Igharghar, running from the same mountain massif northward. The Igharghar has been supposed to end in the Grand Erg, the dune region in southern Algeria, but recent explorations have shown that it does not pass through the Hammada Tinghert, an east-westerly mountain range on the 28th parallel (Kilian, 1925, p. 892). At the point where the wadi was supposed to flow there is, to be sure, a wadi, but it runs from north to south.

Along these rivers there were steppes as is indicated by the northward as well as the southward extension of plants and animals. Among animals that migrated northward were the Carthaginian elephant, a distinct species exterminated by the Romans, and the hippopotamus, which also existed during the time of the Romans. Other mammals of Ethiopian affinity still survive in western Morocco. The most significant flora is that of the high parts of the Ahaggar. Of its 192 sub-alpine species 9 are cosmopolitan or cultivated, 9 endemic, 62 Mediterranean, 11 Ethiopian, and 101 Saharan. One of the endemic forms is a species of *Nananthea*, a genus composed of but two species the other of which occurs only in Corsica. Most of the Mediterranean plants do not occur on intermediate stations between the Mediterranean coast and the Ahaggar.

Calcareous tufas on the border of the eastern scarp of Kharga Oasis in the Lybian Desert indicate a time of greater rainfall, and the presence in the tufas of leaves of an oak (*Quercus ilex*) and other plants which do not flourish now south of Corsica and southern France show that the climate of that period was several degrees colder than the present (Hume and Craig, 1911; Krenkel, 1925, p. 162). Large gravel terraces, especially at Kena (26 ¼° N.) on the Nile, and old wadis descending to the Nile from the Red Sea mountain range show that the precipitation was heaviest where the range is highest and decreased towards the north where the mountains are lower. The rainfall moreover was most abundant on the western slopes (Hume and Craig, 1911; Gautier, 1923, p. 54).

Thus there are distinct traces of greater moisture during the

Pleistocene in nearly all parts of the Sahara and especially in the Ahaggar Mountains, the Atlas, the ranges on the Red Sea, and also in regions crossed by rivers that rise in the above-named mountains.

However, the humidity in northern Africa during the Pleistocene was not too pronounced. The region of the shotts southeast of Biskra was closed; its water did not overflow the sill separating it from the adjacent Mediterranean (Gautier, 1923). Not all of the north African desert enjoyed a moister climate and formed steppes. The Lybian Desert, between the Nile and the sand-hill region, in its central parts lacks all traces of old wadis (Gautier, 1923, p. 52). The cause of this is that the rivers coming from the Red Sea mountains were captured by the Nile and their waters diverted to the Mediterranean. Immense areas were deserts during the pluvial epoch as well as today.

In several cases the postglacial and modern desiccation of rivers and the increased aridity of Sahara are due to physiographic, not climatic, causes; and so the climatic contrast between the pluvial and the present is less than the actual conditions might indicate. Among factors of particular importance have been capture of rivers, the clogging of river valleys by sand dunes so that the former watercourse below the dunes has become a lifeless desert, and tectonic movements that altered the slope of valleys so as to prevent the flow of rivers (Chudeau, 1918, p. 81; 1921, p. 606). Rivers whose sources were in the rainy regions to the south of the desert, particularly, have been subject to piracy. The Niger, which originally emptied into the vast basin 600 km. (370 miles) north of Timbuktu was captured by the Tafassasset probably towards the end of the pluvial period and diverted from the desert to the Gulf of Guinea (Chudeau, 1919).

Another important river damming and capture is that of the Shari, the river feeding Lake Chad (Gautier, 1923, p. 58). Lake Chad is an extensive shallow body of water with numerous islands, shifting outline, and extraordinary variations in extent according to the quantity of water supplied by the Shari. The lake lacks visible outlet but nevertheless has fresh water. This

fact has led many travelers to assume the existence of a subterranean outlet. Attention has been especially directed to Bahr-el-Ghazal, the valley prolonging the lake basin from its southeastern corner towards the northeast. This dry valley, 700 km. (435 miles) northeast of Chad, ends in an immense basin about 100 m. (330 feet) below the level of Lake Chad. The basin on the north and east is bordered by high mountains, while on the south and west a vast lowland extends all the way to Lake Chad. In this basin extensive lakes have existed during recent times. This is suggested by the topography as well as by a rich subfossil mollusc and fish fauna composed of the same species as still live in Lake Chad. The Shari River has been shortened 700 km. (435 miles), because it has obstructed its own course by deposits. To some extent the shortening is due to overflow from a tributary of the Shari to a tributary of the Benué, which is a branch of the Niger, that temporarily, during high water, takes place in a swampy region permitting communication between the two river systems. This is apparently piracy in its first stage. However, there is, Gautier thinks, little doubt that active capture also exists and has played part in the decrease in the quantity of water supplied to Lake Chad by the Shari.

On the contrary, Lake Chad and its fresh water is regarded by A. Penck (1914, p. 288) and Berg (1913, p. 6), as evidence of increased moisture of the southern borderland of the desert belt in modern time.

The fossil ergs or vegetation-covered dunes south of the modern desert, between the Senegal River and Lake Chad, lie largely in depressions with pans which may be chiefly the remains of old river beds (Jaeger, 1926, p. 63). As these may be abandoned simply because of changed watercourses, it does not necessarily follow, according to Jaeger, that the region during the pluvial age formed a desert and later was changed to steppe as held by Chudeau (1921a, p. 605).

At present the Sudan receives rain in summer when the high temperature of the Sahara causes the development of a low here which permits moist southwest winds from the Gulf of Guinea to

prevail as far as to the southern boundary of the desert (Kendrew, 1922, p. 29). Winter is practically rainless, because the dry northeast trades then sweep nearly down to the Guinea coast. The aridity during the pluvial epoch must have been due to dry trade winds throughout the year and to absence of rain-bringing southwest winds in summer.

The fine red muds brought from the Abyssinian highlands by the Sobat, the Blue Nile, and the Atbara, which at present account for 96 per cent of the flood proper of the Nile, to the south of Cairo are at most 9 to 11 m. (30 to 35 feet) in thickness (Hume and Craig, 1911). Ten feet (3 m.) of this have been laid down since the time of Ramses II, that is since 1250 B. C. "If conditions have remained uniform, this would give a date fourteen thousand years ago for the first deposits of alluvial muds in Egypt. Previous to this the mud-laden waters of the Abyssinian Nile system did not reach Egypt, as the waters of Khor Gash now fail to reach the Nile." Thus also in Abyssinia the rainfall during the pluvial period was much less than at present. As a consequence Abyssinia seems to have lacked Pleistocene glaciation, though a number of ranges exceed 3000 m. (10,000 feet) and one peak reaches 4620 m. (15,160 feet). At present the summits often carry snow caps for a long time (Krenkel, 1925, p. 199).

In our time Abyssinia receives heavy rains from monsoon winds in summer, probably owing to the development of a low-pressure system over the upper and middle Nile Valley as the heat increases in spring (Kendrew, 1922, p. 43). Abyssinia lies near the northern border of the region affected by monsoon rains; and during the pluvial epoch, when the temperature was lower, these rains apparently failed to reach so far north.

### Asia

The present deficient rainfall in southwestern Asia and in the region east of the Caspian Sea is caused by constant northerly winds which are dry because of the vast expanse of land and cold seas to the north and because they move towards warmer regions. The aridity of central Asia is of course due to its great distance

from the sea, to precipitation of the moisture of the summer monsoon in the high mountain ranges to the south, and to great evaporation.

At the climax of the last glaciation and the commencement of ice retreat the greater part of the arid and semi-arid zone in Asia may have had a lower temperature and a more humid climate, for mountain glaciation was extensive (see pp. 4–6); and it is in all probability to this age that the high level of a number of lakes belongs. The climate of Asia Minor during the last glaciation is too little known (Frey, 1925, p. 214); but that of Syria and Palestine was cooler and moister with higher lake levels (Blanckenhorn, 1921, p. 10). So it was also in Armenia, where many lakes stood higher and bodies of water occupied basins now dry (Oswald, 1912, p. 18).

"Au-dessus de la mer Caspienne il y a des terrasses à 80–85 m. plus haut que son niveau actuel, auxquelles correspondent celles qui dominent de 4 m. la mer l'Aral. Au-dessus du niveau actuel du Balkache s'élèvent des terrasses jusqu'à 30 m. On a trouvé sur l'Issyk-koul (42 ½° N., 77° E.) des terresses à une grande hauteur. Le Koukou-nor (37° N., 100° E.) en a à la hauteur de 25–50 m. Enfin, le niveau des lacs du Tibet était beaucoup plus élevé (tous les lacs que je vais nommer sont des lacs fermés); ainsi, auprès du Lakkor-tso (32° N., 84° E.) l'on a observé des terrasses à 133 m. au-dessus de son niveau actuel. Au-dessus du grand lac Selling-tso sous le 89° de lat. N., situé à l'altitude d'environ 4900 m., il y a des terrasses qui le dominent de 50 m., et d'autres de la même hauteur au-dessus du lac Panggong-tso (33 ¾° N., 79° E.) sur la frontière entre le Tibet et le Kashmir. Il y en a aussi au-dessus du Manasarovar (30 ¾° N., 89° E.)" (Berg, 1913, p. 19). However, the highest terraces may not date from the last glaciation. The numerous small, strongly saline lakes on the steppes north of the Caspian Sea are remnants of once larger bodies of water probably of pluvial age (A. Penck, 1914, p. 290). The floral exchange between the arctic regions and the Altai that took place during the glacial epoch proves a then moister climate, and so does the modern distribution of

*Artemisia rupestris* in western Siberia and southeastern Russia (Kulczyński, 1924, pp. 137, 166). Lake Baikal during the Quaternary seems to have stood more than 1000 m. (3280 feet) higher (Johansen, 1925, p. 30).

In Persia no sure traces of glaciation have been found, but in the eastern part vast lacustrine deposits and abandoned shore lines tell of earlier greater moisture (Huntington, 1905; Hedin, 1910; Stahl, 1911, p. 21). In the deserts of Sind in northwestern India there are clays of extensive distribution in large tracts covered over by sand (Cotter, 1923, pp. 205, 209, 210). At the present time the low-lying parts, where the impervious clay has remained uncovered by sand and where the dunes form a network, are occupied by lakes frequently of high salinity. The clay, which evidently required much moister conditions for its formation than now exist, may date either from a time of greater rainfall in the region or a time of greater flood of the Indus River due to heavier precipitation in the Himalayas or due to the melting of snow fields and glaciers there.

In the mountain regions west of Peking, China, so far as known a steppe climate similar to the present prevailed almost throughout the Quaternary period (J. G. Andersson, 1925, p. 317). Only after the formation of the loess deposit overlying a mousterian culture, i. e. during or after the climax of the last glaciation, was there a break in the semi-aridity indicated by the deep erosion of the rivers. As this moist period probably was also mild and as the high Wutai-shan Mountains (2890 m.; 9480 feet) in the same district lack traces of Pleistocene glaciation (Olbricht, 1923, p. 727), aridity and loess formation may have prevailed in these regions during the glacial-pluvial epoch. The epoch of heavier rainfall then came in postglacial time.

## Southwestern North America

The Great Basin has deficient rainfall at the present time, because summer temperature is high and evaporation consequently great, because mountain barriers deprive the westerlies of their moisture, and because the winds on descending the eastern

slopes become warmer so that their capacity for holding moisture is increased. The Mexican plateau is arid also because the surrounding mountain rim causes the vapor to be condensed. The west coast between 33 and 28 degrees of latitude suffers drought, since the constant northwest winds, coming from the subtropical high-pressure belt and moving southward, bring little rain.

At the climax of the last glaciation the greater part of these desert regions had a much moister climate than now (Russell, 1885; Gilbert, 1890; Meinzer, 1922). Bodies of water covered large areas, particularly in the northern part of the basin. In the frontier districts of the United States and Mexico the climate was but slightly moister than in modern time. The greater humidity was due both to less evaporation because of lower summer temperature and to greater rainfall than at present (Antevs, 1925a, p. 73). The heavier rainfall is attributed by Huntington and Visher (1922, p. 126) to more frequent and more slowly moving storms.

In southern Mexico, on the southern border of the arid belt, Ixtaccihuatl (5280 m.; 17,320 feet) during the Pleistocene was glaciated down to the level of 3400 m. (11,155 feet) or 1180 m. (3860 feet) below the modern glacier line, and Popocatepetl (5440 m.; 17,850 feet) to below that of 4000 m. (13,125 feet) (Jaeger, 1925, p. 372; 1926). Lake Mexico, which now is artificially drained, once reached, as terraces show, 53 m. (174 feet) above the valley floor (Jaeger, 1926, p. 38). No evidence was found of interruption of the moist climatic conditions by aridity. The Pleistocene age of the high level of the lake is proved by the occurrence in the deposits of bones of *Elephas imperator*. Lake deposits occur down to the southernmost part of the Mexican plateau. Jaeger concludes that the arid belt of North America during the Pleistocene retreated on the equatorial side as well as on the polar side.

## Southwest Africa

In the arid regions of the southern hemisphere conditions similar to those north of the equator prevailed.

The present aridity of southwest Africa is due to the prevailing southeast trades, which become dry on crossing the high southeastern part of the continent, and to the cold Benguella Current which cools off the overlying air so that sea winds blowing in over the warm continent soon acquire greater capacity of moisture and become dry. In winter a subtropical high-pressure area rests over the southeastern part; and this season is dry except in the southwest of the Cape Town province which is skirted by the northern fringe of the westerlies.

In the Great Karroo (33° S.), in the southern part of the South African Desert zone, imperfectly developed pans and great thickness of alluvial deposits and, in the southern part of the Kalahari Desert (28° to 26° S.), river channels now occupied by intermittent rivers point to once greater humidity. These conditions in Passarge's opinion (1904, pp. 648–668; 1904a) indicate that the region had a marked pluvial period. The complete absence of Pleistocene glaciers in South Africa, even on Mont Aux Sources (28° 45′ S., 29° E.) which is about 3350 m. (11,000 feet) high, makes Rogers (1922) inclined to believe that the precipitation was little heavier than today but that the lowered temperature, which according to the general view must have existed, caused the rainy belt to extend farther north than now and caused an increase of flowing water in what are now wadis, so that possibly some of them held permanent rivers for a long period. The actual difference in opinion between Passarge and Rogers seems to be much less than Rogers' wording would lead us to suppose. The precipitation on Mont Aux Sources may hardly have been influenced by the equatorward migration of the rain-bringing westerlies; and so the absence of glaciers there may have no bearing on the rainfall in the dry area, which seems to have been essentially heavier than now.

At the same time the dry belt may have lain farther north over the sand regions of the northern Kalahari. This is suggested by the occurrence of a modern vegetative covering on old dune regions in the northern Kalahari, south of the 20th parallel, and by the occurrence of pans in the Ermelo district (26½° S.,

30½° E.) in eastern Transvaal and near Heidelberg (26½° S., 29° E.) in southern Transvaal, that is in regions now much too moist for the development of pans.

The Etosha Pan (18° 50′ S., 16° 20′ E.) gives no clue as to the climate, as its recent change from a permanent lake to a pan was at least chiefly due to diversion of its tributary to the ocean (Jaeger, 1926, p. 64). In the Namib Desert (19° S., 13° E.) arid conditions and the formation of pans date from the Miocene (Kaiser, 1922, p. 164).

The present conditions were in existence long before records were made by the white man.

## Australia and New Zealand

The modern arid climate of Australia is due to the fact that the continent lies in the calms of the subtropical high pressures and the trade winds which blow out from them towards the equator. The rain-bringing Antarctic cyclones skirt the south coast only in winter. The interior, because of the permanent high pressure, has at best scanty rainfall. The north receives rain from monsoon winds in summer, when the high-pressure belt has migrated far to the south. The climatic conditions during the Pleistocene have been well summarized by the Climate Committee of the Australasian Association for the Advancement of Science (Speight, 1914, 1921a). The paleontologic and physiographic evidence shows that at the close of the Tertiary the climate grew colder and that the refrigeration was attended and followed by a more pluvial climate, which in its turn was succeeded by drier conditions continuing up to the present.

First among the evidence of cooler and moister climatic conditions in Pleistocene time comes, of course, the previous larger extent of the glaciers (see p. 10). In southeastern Australia the former heavier rainfall is evidenced by enormous accumulations of estuarine deposits, by recent clogging of river channels, by long lines of *Eucalyptus populifolia* and allied types indicating abandoned river courses, by widespread remains of *Diprotodon australis*, an extinct giant marsupial, in present desert regions, etc.

(Speight, 1914, pp. 244, 245; 1921a, pp. 343, 349; G. Taylor, 1919). Diprotodon remains occur as far north as southwestern and central Queensland. Eucalypts and acacias of the cold plateaus of New South Wales tell of northward migration enforced by cooler climate and an inability to migrate from certain isolated plateaus after the passing of the cold period (Speight, 1914, p. 246). In Tasmania the flora, by absence of many types occurring on the mainland, shows effects of former refrigeration.

In the deserts of Western Australia old deep river valleys indicating strong water flow have subsequently been clogged as the streams, evidently because of increasing aridity, have become incompetent to carry their load (Speight, 1921a, p. 349).

In the northern part of Australia, on the contrary, the climate previously was drier than at present. In Northern Territory the former aridity is evident from the large extent of laterites and porcellanites which cap various geological formations. Today the laterites are undergoing rapid denudation by rivers. The first indication of the return to a more humid climate was the formation of bog-iron ore in depressions. Also in Cape York Peninsula and other parts of northern Queensland laterites and porcellanites indicate a former drier climate, as does the present capture of the gulf rivers by the coastal rivers emptying into the Pacific (Speight, 1921a, pp. 347, 348).

The moister climatic conditions during the Pleistocene apparently were due to a temperature fall through which the rain belt lay farther north or over southern Australia and the arid belt was displaced to the northern part of the continent.

In the South Island of New Zealand the extensive Pleistocene glaciation (see p. 10) indicates that the climatic conditions then were different from those that prevail now. Today the island lies in the variable westerlies all the year. The rainfall is excessive on the west coast and in the mountains but decreases rapidly east of the mountain range. The condensation of moisture in the mountains liberates latent heat, and the rain-bringing winds as they descend the eastern slopes become abnormally warm and dry and cause considerable evaporation.

The chief difference during the glacial age may have been that the temperature was lower and the land stood higher in relation to sea level. The precipitation on the west coast and in the mountains may have been about as today. In the eastern part of the island it may have been less, because of the greater height of the mountains in the west, and here pronounced steppe conditions are believed to have prevailed at this time, as is suggested among other things by the distribution of *Ranunculus paucifolius* (Speight, 1907, p. 31; 1914, p. 246; 1921a, pp. 350, 353, 354). However, if the origin of the New Zealand flora as a whole is considered, "it is obvious that there can have been no marked refrigeration of the climate since Tertiary times, since that would have resulted in the extirpation of the Malayan tropical and subtropical element which, even under present conditions, is only just maintaining its ground against the Antarctic element, unless the distribution of the land was such that it allowed the Malayan plants to migrate northward before an advancing ice sheet, and to come south again as the climate grew milder. There are, however, serious obstacles in the way of this explanation" (Speight, 1914, p. 249).

When the glaciers retreated the climate is believed to have become moister and milder than it is now. This is shown by the occurrence of stumps and trunks belonging to a former totara[1] forest buried under the gravels of the Canterbury plains near Riccarton, for under the present climatic conditions the totara and associated trees could not establish themselves in the region. A former warmer climate is also indicated by the occurrence of certain plants and perhaps also animals in isolated places to the south of their general range of distribution (Marshall, 1912, p. 53). Subsequently there followed a moderate steppe climate which still persists.

## South America

In the border regions of Chile, Bolivia, and Argentina and extending across the entire Cordillera there lies the Atacama Desert, a barrier for the spreading of plants and a frontier of

[1] *Podocarpus totara.*

vegetative regions (Herzog, 1923, p. 231). The desert conditions are due to the dry southeast trade winds and to higher temperature of the land than of the sea because of abnormally cold water on the coast. Previously, however, a floral exchange of Antarctic and North American forms took place across the desert. Several Antarctic plants were more widely distributed in the present desert, as is shown by the disrupted occurrence of a rich vegetation of *Gunnera pilosa* on banks of brooks. The basins like that of Lake Poopo contained larger bodies of water than today (W. Penck, 1920, p. 251; Ogilvie, 1922, pp. 43–45). The climate of that time must have been much moister, even though the Atacama Desert was still largely a semi-desert (Klute, 1925). At the same time the temperature was lower, for in Peru high-Andine plants growing in peat bogs and other moist and cold localities at levels down to 2800 m. (9186 feet) appear to be relicts of the nival flora which in postglacial time has retired to greater altitudes (Weberbauer, 1911, p. 316). Former cooler and moister climate is also shown by earlier more extensive glaciation in the mountains (see p. 9; also Troll, 1927).

The flora and fauna of the Andes from Colombia to Argentina present a remarkable relationship to those of high latitudes in North America and Europe (Bates, 1891; Meyer, 1907, p. 463; Weberbauer, 1911, p. 315; Scharff, 1912; Heintze, 1918; Herzog, 1923, p. 230). The plants show a large amount of identity as to genera but not as to species with North American and European forms. They occur in the alpine regions and form a percentage that increases with altitude so as finally to present about one-half as many genera as the other floristic elements. The faunistic similarity is particularly striking as regards insects, and the percentage of animals of northern affinity increases with elevation. Numerous genera which are totally absent from the intervening tropical and subtropical zones of America are common to the three continents. The relationship has no reference to species. The migration of both the plants and the animals from the northern to the southern continent must have taken place during times of lower temperature than at present, for only then could they

have been able to cross the gaps in the mountain chains in Central America and the now tropical Panama region. The migrations evidently occurred during the glacial-pluvial epochs. The insects especially are by some scientists believed to have reached South America long before the last glacial period. When subsequently the temperature rose the biota were driven up in the mountains, and connection with their original areas of distribution was cut off. Early immigrants, of course, moved down and up the mountains, as the glaciers later expanded and retreated.

In southern Brazil also there are indications of changes in precipitation during the Pleistocene (Woodworth, 1912, p. 111). In São Paulo (22° S.) and Paranâ (24° S.) there are along several of the larger rivers "terraces of sand and gravel evidently remnants of former aggraded floor of their valleys." On the top of the pebble beds there occurs a red clayey earth, a *terra rossa*. Deposits from six to ten feet in thickness were observed by Woodworth, but in many cases the substratum was not exposed. The deposits occur "particularly along the lower slopes of the hills as if washed down by rains during the wet season. Much dust is blown about by the winds in the dry season, and doubtless an eolian origin may be attributed to some of the particles." Dr. O. A. Derby expressed the opinion that the red earth was an equivalent of the loess of other regions. The dating of the beds offers difficulties, because it is not evident whether a compression of the climatic belts towards the equator would cause a change in the precipitation in the region, or, if so, in which direction. The district at present receives its rains chiefly in summer when the equatorial low-pressure system is farthest south. In winter the region lies in the subtropical high-pressure belt. The gravels probably date from the pluvial epoch; and the red earth, representing drier conditions, may be of postglacial age. But it is also possible that the gravels are interglacial and that the red earth represents the arid zone during the pluvial epoch.

## EQUATORIAL BELT

### Africa

The climatic conditions in equatorial Africa during the Pleistocene are indicated in the first place by the former greater extent of the glaciers on the highest mountain peaks, viz. Kenya, Kilima Njaro, and Ruwenzori. So far traces of only one glaciation have been recognized (see Krenkel, 1925, p. 246). Regarding Mt. Elgon (1° N.; 14,150 feet; 4313 m.) no direct statement has been found by the writer. At present it lacks permanent snow and ice fields, the snow that falls not staying long (H. Johnston, 1902, p. 61). Mt. Kenya (5194 m.; 17,040 feet), whose summit lies 13 km. (8 miles) south of the equator, was once covered by an ice cap extending 1524 m. (5000 feet) below the existing glaciers or down to an altitude of about 3050 m. (10,000 feet) (J. W. Gregory, 1894; 1921, p. 149). On Kilima Njaro (3° S.; 5890 m.; 19,320 feet) the lowest level of glacial traces on the southern slopes seems to lie at (3600 to) 3800 m. (11,810 to 12,070 feet) altitude or as much as 1650 m. (5410 feet) below the modern glacier limit (Klute, 1920, p. 131; 1925). On Ruwenzori (20° N.; 5125 m. 16,815 feet) the Pleistocene ice cap perhaps expanded even down to 1525 m. (5000 feet) above sea level, while the existing glaciers end at 4115 m. to 4070 m. (13,500 to 13,350 feet) (Elliot, 1895, p. 310; Roccati, 1907; De Filippi, 1908, p. 391). According to Meyer (1907, p. 471) the glaciers only reached the level of 3100 m. (10,200 feet).

The Pleistocene glaciers occurred in the same situations as the modern ones. On Kenya and Kilima Njaro they extended lowest on the southwest side, on Ruwenzori on the east and southeast sides. The situation on Kenya and Kilima Njaro is believed to be due to the dry anti-trades from the northeast. As for Kenya it may also be important that the rainfall, which chiefly takes place when the sun is overhead and the winds are weak and variable, is much heavier on the southern side than on the northern (Kendrew, 1922). This greater extent of the glaciers shows that the climate was either cooler or moister or, most likely, both.

Abandoned shore lines marking higher levels of the lakes and shore lines and deposits indicating earlier existence of water in now dry basins of the steppes also tell of a former moister climate (Tornau, 1913, p. 45; Hobley, 1914; Harger, 1917; Brooks, 1925, p. 104; Krenkel, 1925, pp. 266, 271, 320, 326, 338). So do also the large and numerous caves in the basalt tufa on Mt. Elgon, which probably have been formed by running subterranean water during a time of greater rainfall (Lindblom, 1921, p. 157).

Evidence to the same effect is furnished by flora and fauna. On the high mountains from Abyssinia to Kilima Njaro, at altitudes between 1980 and 3050 m. (6500 and 10,000 feet), there is a more or less common flora which recalls trees and plants of temperate South Africa, of the Mediterranean, and of boreal Europe and which is quite unlike the vegetation in the intervening lower land (Meyer, 1900, pp. 398–403; H. Johnston, 1902, p. 318; Engler, 1910, p. 989; Hobley, 1914, p. 475). It is characterized by short grass, yew, juniper, Habenaria, giant lobelias, giant groundsels, tree ferns, violets, buttercups, clover, forget-me-nots, geraniums, etc. Also the fauna of the higher regions, especially the butterflies, shows palearctic affinity (Sjöstedt, 1910, p. 57). The spreading of this flora and fauna across the plains that now are steppes proves that the biota once occurred at much lower levels than today, which was possible only during a time of essentially lower temperature and perhaps greater precipitation than at present. As the climate became warmer and drier they withdrew up on the high mountains and disappeared from the intervening land. The date of this isolation lies far back, for in some cases different species have developed on different mountains (J. W. Gregory, 1921, p. 151).

The Quaternary formations on the east coast which show no sign of refrigeration but rather suggest higher temperature (J. W. Gregory, 1921, p. 150; Krenkel, 1925, p. 329), because of the lower sea level during the glaciation, must be of interglacial age, unless the land has undergone great changes of level.

Everything considered, it seems probable both that the temperature was lower and that the rainfall was heavier than at

present. The higher level of the lakes may have been partly due to lessened evaporation as a consequence of the lower temperature.

### South America

In equatorial South America evidence of former greater extent of the glaciers (see p. 8) and floristic and faunistic peculiarities also indicate former cooler and moister climate. The coastal region of northwestern Peru has presumably been a desert throughout the Pleistocene (Bosworth, 1922, p. 241). The aridity is due to the low temperature of the sea on the coast. The marine fauna has remained unchanged during the Quaternary.

### India

In India south of the Himalayas no traces of Pleistocene glaciation have been found, but in various mountain ranges in southern India, in Assam, and on isolated hills such as Parasnath (24° N., 86° E.) and Mt. Abu (24½° N., 73° E.) there occur plants and animals identical with or closely allied to forms existing in the Himalayas (Pilgrim, 1910). "These occurrences of Himalayan species on hills south of the Himalayas and their entire absence in the intervening area is undoubtedly easiest explained by a cold period having affected the plains of India, but it seems possible that it may be due to changes having taken place in the rainfall and in the distribution of forest."

### East Indian Archipelago

In the East Indian Archipelago various conditions indicate heavier rainfall during the Pleistocene, viz. "(a) a rise of the level of the lakes, e. g. Lake Toba in North Sumatra; (b) an increase of the accumulative power of the rivers which enabled them to depose more gravel, which was the cause of the building up of the river terrace e. g. of the Wampoo in North Sumatra; (c) a more effective weathering of the rocks by which, in my opinion, on the isle of Java, the red volcanic soil, often but incorrectly called laterite, was built" (van Baren, 1922, p. 10).

On Mt. Kinabalu, 4100 m. (13,455 feet), in the northeastern

corner of Borneo the dwarf forest at 3200 m. (10,500 feet) of elevation shows relationship with the European-Asiatic-Himalayan region on the one hand and with the Australian-New Zealand-South American on the other (Stapf, 1894). The spreading would have taken place most readily during an age of cooler climate. The peak shows traces of old glaciation (Bowman, 1916, p. 206). Occasionally ice is found on the summit (Moulton, 1915, pp. 162, 164).

In New Guinea, on Mt. Wilhelmina (4½° S., 138½° E.; 4750 m.; 15,580 feet), where perpetual snow, but no real glaciers, now is to be found at 4536 m. (14,883 feet), a lake, striated boulders, and rock ledges have been observed at 4054 m. (13,300 feet) of altitude proving the former existence of real glaciers much below the modern snow fields (Lorentz, 1911, pp. 492, 496, 500). On the northern slope of Mt. Carstensz, New Guinea, which rises to 4785 m. (15,700 feet), i. e. 457 m. (1500 feet) above the modern snow line, there are numerous traces of former glaciation down to the altitude of 3660 m. (12,000 feet) and perhaps even down to that of 2590 m. (8500 feet) (Wollaston, 1914, p. 265). At the western end of the Huon Gulf (7° S., 147° E.) there occur near sea level—in a region where the bed rock for long distances consists of diabase—enormous, rounded, polished, and striated erratics of granite and also walls that cannot be anything but moraines (see Gagel, 1912, pp. 15–18). The finds may be evidence of Pleistocene mountain glaciation, the region probably having undergone a later subsidence of many thousand feet. More exact dating of the glaciations has not been attempted. Thus also in New Guinea the climate during the Pleistocene was more favorable for the development of glaciers.

Mauna Kea, Hawaii, 13,825 feet (4214 m.) high, was capped by ice extending down to 11,500 feet (3505 m.) of altitude and covering an area of some 10 square miles (H. E. Gregory, 1925).

## Climate during Ice Retreat

Some time after the inauguration of the climatic change which checked the growth of the ice sheets wastage began to exceed

supply, and shrinking started. The chief governing factors of ice retreat are high summer temperature, strong insolation, clear sky, and little precipitation in solid form (see Antevs, 1925b, p. 45). When the ice edge ends in water calving also plays a very important rôle; but primarily the rate of recession is an exponent of the summer amount of heat. The measured rates of retreat of the ice border show that during most of this stage the climatic conditions were very favorable for ice wastage and as time passed became increasingly so. Halts and readvances, however, indicate that this was not always the case. But even when the insolation was strong the heat was largely used in fusing the ice and warming the ice-cold melt-water and the ground, so that the air temperature on and in the vicinity of the ice was fairly low. Cold winds sweeping down from the ice sheets also tended to keep the temperature low. In spite of this the climate was not directly comparable to that in modern arctic regions both because of difference in latitude and because in the arctic regions at the present time the glaciers, even if but moderately nourished, hold their position. On south-facing slopes protected from cold winds strong insolation could raise considerably the temperature of the ground and of shallow water bodies.

The flora and fauna that took possession of the uncovered land during the early days of retreat in Europe and North America indicate tundra conditions in a belt bordering the ice. In Germany there lay outside the tundra a steppe belt, and still farther out a forest belt, whose pioneers were birch, pine, and aspen. Both tundra and steppe were essentially semi-arid. These belts moved northward as the ice front receded, and the forest belt gradually gained ground on the other two, so that the steppe belt was eliminated in Denmark and the Dryas belt in central Sweden.

In Europe some of the arctic biota withdrew both to the north and to high regions in the Alps and other mountain ranges, while others only retired to one of these areas. The steppe flora and fauna virtually withdrew to southern Russia and western Asia.

# CHAPTER III

## ALTITUDE AND SLOPE OF THE SURFACE OF THE LAST PLEISTOCENE ICE SHEETS IN NORTH AMERICA AND NORTHERN EUROPE

While the areas occupied by the two greatest Pleistocene ice sheets are distinctly marked out and except in the arctic archipelago fairly well determined, the third dimension—thickness—is indicated obscurely if at all and is very difficult to find out. If we look, on the one hand, to the modern ice sheets of Greenland and the Antarctic, where direct measurements of altitude of the ice surface have been made by many explorers, we are handicapped by lack of knowledge of the altitude of the land beneath them and embarrassed also by the realization that both as regards nourishment and wastage these two almost stationary ice sheets near the poles are very different from those vigorous sheets of ice that spread far out in mid-latitudes. If we seek, on the other hand, by more direct methods to work out the thicknesses of these old ice sheets by locating and comparing marks made at their upper limits, on nunataks and fringing highlands, we find, particularly in North America, that the regions occupied were almost totally buried by the ice. Nunataks were rare and were pretty strictly confined to the marginal zone. So we are obliged to substitute minimum figures of altitude and thickness for exact ones, assuming that the highest glaciated summits of the interior were covered, and no more; or else we must make a pure guess. Other factors of course present difficulties, but they are of less consequence.

Because it is a problem of interest in itself, but more especially because we must have a basis for figuring the volume of ice gathered into these glaciers and subsequently returned to the sea, before we can estimate the fluctuation of sea level caused by this exchange, we shall endeavor in spite of the difficulties just

mentioned to work out in this chapter the probable cross sections of the North American and European ice sheets.

## Conclusions Drawn from the Modern Greenlandic Ice Sheet

The Greenlandic ice sheet is on the whole highest somewhat to the east of the medial line (Atlas of Greenland). The greatest elevations encountered during the various crossings range from 2225 to 2950 m. (7300 to 9680 feet). The last-mentioned altitude was measured about in the center of the ice shield. The eastward inclination of the surface in the interior from the crest of the ice was found by J. P. Koch to be 1 in 777 (from 2950 m. to 2500 m. in 350 km.), by Knud Rasmussen 1 in 666 (from 2225 m. to 2000 m. in 150 km.), and by Nansen (1890, p. 465) 1 in 142. The westward interior slope was found by Koch to be 1 in 347 (from 2950 m. to 2200 m. in 260 km.), by Rasmussen 1 in 337 (from 2225 to 1810 m. in 140 km.), by Alfred de Quervain 1 in 400 (from 2500 m. to 2000 m. in 200 km.), and by Nansen 1 in 206. Koch, as is known, crossed the ice at the center, Rasmussen near the wide northern end, de Quervain somewhat to the south of the widest part, and Nansen in the southern, narrow part. Thus the central parts are by far most even; and the wider the ice sheet, the flatter the surface.

In the marginal belt on the east side the following gradients, calculated from the elevation of the surface of the ice down to sea level, were encountered: 1 in 120 (Koch; 2500 m. elevation 300 km. inland), 1 in 90 (de Quervain; 2500 m. 220 km. inland), and 1 in 75 (Rasmussen; 2000 m. 150 km. inland). And on the west side: 1 in 57 (Koch; 2200 m. 125 km. inland), 1 in 100 (de Quervain; 2000 m. 200 km. inland), and 1 in 90 (Rasmussen; 1540 m. 140 km. inland). Along Nansen's (1890, p. 465) route the gradient on the east coast for the ascent from sea level to 1000 m. elevation was 1 in 23, and for that from 1000 m. to 2000 m. 1 in 57; on the west side the corresponding gradients were 1 in 42 and 1 in 84. In the marginal belt of the northeastern corner Mikkelsen (1922, Pl. 1) found a highest altitude of 1050 m.

about 60 km. from the coast which gives a gradient from sea level of 1 in 60. The altitudinal curves on Mikkelsen's map shows that the slope varies considerably. At Danmarks Fjord the ice surface falls from 700 m. to sea level in 10 km., or with a gradient of 1 in 14. Other calculations of the surface slope have been made by Meinardus (1926, p. 103).

Thus the Greenlandic ice sheet has the shape of a very gently arched dome with steep margins. From the nearly flat interior the surface falls with gradually increasing gradient and finally quite abruptly towards the margins. In northern Greenland the edge normally forms a high precipitous bluff which only locally can be ascended. In southern Greenland, where temperature is higher and surface ablation is greater, it normally forms a slope—only exceptionally a vertical wall.

The Greenlandic ice sheet is now, generally speaking, stationary and has not undergone any noteworthy increase or decrease during the last 1000 years (Hammer, 1921, pp. 7, 9). To be sure, the ice border in historic time has retired at some places and advanced at others, but on the whole it has not moved much since the Northmen colonized Greenland in the tenth century, for several ruins dating from that time occur close to the ice edge. This means that supply and wastage are about the same.

On the northern side of the Great Karajak Glacier on the west coast of central Greenland the annual ablation at the altitude of 200 m. (655 feet) may amount to 2.4 to 2.8 m. (7.9 to 9.2 feet) and somewhat to the west (70° 35′ N., 52° W.) at elevations of 100 to 570 m. (330 to 1870 feet) to 2.00 to 2.25 m. (6.6 to 7.4 feet) (von Drygalski, 1897, pp. 253, 338, 354, 357). The mean annual amount of ablation of the Greenlandic ice sheet in a marginal zone 10 km. (6 miles) wide is estimated by Hess (1904, p. 245) at 1 m. (3.28 feet). Where the ice ends on land and is stationary the forward motion consequently amounts to a meter or a few meters annually. The rate of motion of the ice sheet somewhat inside the beginning of the Great Karajak Glacier, was so small near land as to be imperceptible, and at a distance of one to a few kilometers out was 0.3 to 0.4 m. (0.98 to 1.3 foot)

in 24 hours (von Drygalski, 1897, p. 220). At another place on the northern side of the same glacier and 250 m. (820 feet) from land the ice movement amounted to 2 cm. (0.79 inch) a day (*loc. cit.*, p. 239). The average motion of the ice sheet is estimated by Chamberlin and Salisbury (1905, p. 261) at less than one foot (0.3 m.) a week.

Where ice descends to the sea as glaciers in the fiords and as an integral part of the ice sheet and where it then eventually ends in a floating front, the motion is much faster. The largest glaciers move at rates of 15 to 30 m. (50 to 100 feet) in 24 hours and at about an equal rate at all seasons, in other words move 5.5 to 11 km. (3.4 to 6.8 miles) a year and discharge great numbers of large icebergs (Hammer, 1921, pp. 10, 13). However, the great majority of the glaciers are not very productive, and some are even quite stationary. The surface slope of the glaciers is mostly just a few degrees—5° approximately corresponds to a slope of 1 in 12½—but frequently as much as 10° to 15° and in some cases even more.

Thus both the wastage and the nourishment of the Greenlandic ice are insignificant in comparison with the changes in those parts of the Pleistocene ice sheets that extended down into low latitudes. Consequently the shape of the Greenlandic continental ice cannot be taken as a norm (cf. p. 72).

## Conclusions Drawn from the Modern Antarctic Ice Sheet

South Victoria Land and the land to the south are occupied by a vast plateau which is 700 feet (2135 m.) high at its northern end, or at the South Magnetic Pole, and probably 12,000 feet (3660 m.) at the Geographic Pole (David and Priestley, 1909, p. 303; Amundsen, 1912; G. Taylor, 1913, p. 287). The surface of this plateau consists of névé, and beneath the firn no doubt is glacier ice. In the interior the plateau is quite flat. Along Amundsen's (1912, Vol. 2, map) route the ice shield fell from 12,260 feet (3735 m.) at the pole to 9730 feet (2966 m.) on the 86th parallel, or in 210 miles (338 km.), that is with a gradient of 1 in 438 or 12 feet to a mile. According to Scott the altitude at the pole is

9070 feet (2765 m.) and at the 89th, 88th and 87th parallels respectively 9630, 9810, and 9392 feet (2935, 2990, and 2862 m.) (Wright and Priestley, 1922, Pl. 3). In the part called King Edward VII Plateau the gradients along Shackleton's (1909, map) route are in outward direction 1 in 580 (from 10,050 feet elevation to 9050 feet in 110 miles), 1 in 290 (from 9060 feet to 8060 feet in 55 miles), 1 in 387 (from 8050 feet to 7750 feet in 22 miles), and finally 1 in 128 (from 7750 feet to 6800 feet in 23 miles), after which glaciers begin to descend with steep gradients between mountains on the border of the plateau. Along the line from the South Magnetic Pole to Terra Nova Bay the névé surface at first is rising with a gradient of 1 in 2500 (from 7250 feet elevation to 7350 feet in 48 miles), then falling with a slope of 1 in 980 (from 7350 feet to 7000 feet in 65 miles), and in the outer 120 miles with 1 in 90 (Shackleton, 1909, map). In the interior of Adélie Land and King George V Land the gradient is 1 in 462 (surface at 6000 feet 300 miles inland and at 4000 feet 125 miles inland) (Mawson, 1914, Vol. 2, p. 293). In the marginal belt of the ice in Adélie Land the gradient is 1 in 53 (4000 feet elevation 40 miles inland), and in that of Kaiser Wilhelm II Land 1 in 44 (3000 feet 25 miles from the sea) (Mawson, 1914, maps). Near the margin the slope averages 1 in 70 from 12 to 5½ miles from the coast (elevations 2000 and 1500 feet respectively) and 1 in 19 from 5½ miles inland to the sea (Mawson, 1914, Vol. 1, p. 109). Other calculations of the slope have been made by Meinardus (1926, p. 103). Thus from the flat interior the slope of the surface increases gradually towards the edge of the ice, in the vicinity of which it is quite steep.

The thickness of the ice is possibly about 5000 feet (1525 m.) (David, 1914, p. 613).

The wastage of the ice chiefly takes place through discharge of icebergs. Thaw occurs only locally and occasionally even during midsummer; and ablation consists chiefly in evaporation and wind erosion (Wright and Priestley, 1922).

The relation between nourishment and ablation of the Antarctic ice sheet may be positive, judging from the outflow of ice that

has been observed at most places (von Drygalski, 1919, p. 21); but the relation between the supply and total wastage may be negative, and the Antarctic may now be undergoing deglaciation (Wright and Priestley, 1922, p. 450). The motion of the ice sheet itself is perhaps limited to the marginal belt, but that this part moves is evident from the discharge of icebergs and the pressure exerted on the shelf ice lying outside. The rate of movement of the ice sheet on land seems to be unknown but is said by von Drygalski to be much less than in the sea. It seems frequently to be unnoticeable. The precipitous bluff of the continental ice, where it ended in the sea at Gaussberg, moved about 0.4 m. (1.3 foot) a day (Drygalski, 1906, p. 62). Antarctic glaciers have been found to move at rates of from a few feet to a few hundred feet a year (see Drygalski, 1919, p. 21; Wright and Priestley, 1922, p. 133); and the Ross Ice Barrier, which is largely afloat, has been found to move at the rate of 12 feet (3.6 m.) a day or 4400 feet (1340 m.) a year (see C. S. Wright, 1925, pp. 200, 203, 216, 219).

The Antarctic ice sheet is still less comparable to the Pleistocene ice sheets than is the Greenlandic ice.

## Altitude of the Ice Surface in the Peripheral Belt

### NORTH AMERICAN ICE SHEET

Data on the elevation of the surface of the last North American ice sheet are at hand only from the outer regions of the glaciated area, since in the central parts no mountains reach sufficient heights to furnish information.

At or just inside the terminal moraine in the eastern states the altitude of the glacier surface on present low land was several hundred feet, as the moraine here and there, for instance at Harbor Hill, Long Island, rises at the sea to about 400 feet (120 m.) altitude. The elevation may even have been 1000 feet (300 m.), since a glacier near its front frequently is twice as high as the moraines it forms. On higher land the ice surface reached greater elevations, even if the thickness was not greater. In the high-

lands near the ice border in New Jersey drift is to be found up to about 900 feet (275 m.) above the sea (Smock, 1883, p. 342). On Pokono Knob (41° 5′ N., 75° 25′ W.), near the eastern border of Pennsylvania and just a few miles inside the terminal moraine, the upper glacial limit may lie at about 2000 feet (610 m.), whereas the mountain is 2025 feet (617 m.) high (Smock, 1883). In the Moosic Highlands in the northeastern corner of Pennsylvania, about 60 miles (100 km.) inside the outermost morainic line, it lay above 2700 feet (825 m.), glacial striae having been observed on the summit of the highest peak, Elk Mountain (41° 43′ N., 75° 40′ W.) (Branner, 1886, p. 365). To the east of these mountains Walnut Hill, or Liberty Hill, in New York exhibits glacial erosion and drift to an altitude of 2000 feet (610 m.) and towered above the ice shield (Smock, 1883). On the other hand the highest peak in the Shawangunk Range on the same latitude in New York, reaching an elevation of 2340 feet (713 m.), was covered by the ice. On Slide Mountain (42° N., 74° 25′ W.; 4205 feet; 1282 m.), the highest summit of the Catskills, traces of the last ice sheet have been found up to 3900 feet (1190 m.) altitude according to information from Dr. John L. Rich. Here a few peaks consequently formed nunataks.

In southern New England a few minimum figures of the altitude reached by the ice have been obtained. Mt. Everett (2626 feet; 800 m.), in the southwestern corner of Massachusetts, was overtopped by thick ice (Dana, 1875, p. 168). So also were Mt. Wachusett (2108 feet; 642 m.) in central Massachusetts (Alden, 1924, p. 36) and Mt. Monadnock (3186 feet; 971 m.) in southern New Hampshire (E. Hitchcock, 1841, p. 389; Goldthwait, 1925, p. 15). Mt. Greylock (3535 feet; 1077 m.) in the northwestern corner of Massachusetts, the highest peak in the state, was buried, the very top showing glacial striae according to information from Professor H. F. Cleland.

In Vermont the ice completely covered all mountains; their highest summits, Mt. Mansfield and Mt. Killington, are 4406 feet and 4380 feet (1342 and 1335 m.) respectively (C. H. Hitchcock, 1904; Goldthwait, 1915–16, p. 71). Whether it overtopped the

Adirondacks in northern New York is not settled, C. H. Hitchcock (1904, p. 80) being inclined to believe that the loftiest peak, Mt. Marcy, 5344 feet (1629 m.) high, was buried, while Coleman (1920, p. 324) thinks that the ice reached only an altitude of about 4800 feet (1465 m.). The great height to which ice action is recorded in the Catskills suggests that the ice surface reached high above the loftiest summits of the Adirondacks. In northern New Hampshire the ice surface (probably during the last glaciation) rose above the highest mountains in the northeastern states, the White Mountains, which reach 6293 feet (1918 m.) altitude (C. H. Hitchcock, 1876; Goldthwait, 1913, p. 3; 1925, p. 13). In Maine it overtopped the highest peak in the state, Mt. Katahdin (5273 feet; 1607 m.), and Mt. Abraham (3388 feet; 1033 m.) (Tarr, 1900, pp. 435, 436), and the summit of Mt. Desert (1527 feet; 465 m.), the highest point of the coast of the state (G. F. Wright, 1900, p. 166). Thus with exception of a few peaks all mountains in the eastern states well inside the terminal moraine were buried beneath the ice.

On the Gulf of St. Lawrence, at and near the edge of the ice sheet, the ice during its greatest extent reached only to inconsiderable elevations above the present sea level, even leaving islands only a few hundred feet high rising above it (Coleman, 1920, p. 322). In the Magdalen Islands till occurs only up to 105 feet (32 m.) altitude (Goldthwait, 1915). In the southwestern part of Gaspé Peninsula the ice reached up to at least 1270 feet (387 m.) and 100 miles (160 km.) farther northwest to less than 3000 feet (915 m.) altitude (Coleman, 1920, p. 324). In the Shickshock Mountains in the peninsula, which are as much as 4300 feet (1310 m.) high, traces of local glaciers, not of the ice sheet, are found up to 3000 feet (Coleman, 1920, p. 323).

In northeastern Labrador, in the Torngat Mountains (58° to 60° N.) reaching altitudes of over 5000 feet (1525 m.), distinct evidence of glacier action was observed up to 2650 feet (810 m.) (Coleman, 1920, p. 321). On Johannesberg (58° 20′ N.), the ice did not reach within several hundred feet of the flat summit at 2300 feet (700 m.) elevation. On the Mugford Mountains

(58° N.) the surface of the ice attained an elevation of 2000 feet (610 m.).

In Montana (48° N., 107° to 113° W.) the surface of the ice during its greatest extent stood at levels of from 3500 to 4720 feet (1065 to 1440 m.) or at an average height of 4000 feet (1220 m.) (Calhoun, 1906, p. 28). At the intersection of the 49th parallel and the 113th meridian, Alberta-Montana, numerous Laurentian boulders have been found at an elevation of about 4200 feet (1280 m.) (Dawson and McConnell, 1884, p. 148). About 20 miles (32 km.) north of the 49th parallel the same kind of erratics were traced even up to 5280 feet (1610 m.) of altitude. In the Cypress Hills (50½° N., 109° W.), Saskatchewan, glacier action has been observed up to an elevation of 4400 feet (1340 m.) or to within 400 feet (120 m.) of the highest summit, and in the Hand Hills (51° 40′ N., 112½° W.), Alberta, up to that of 3400 feet (1035 m.) (McConnell, 1886, p. 75).

## EUROPEAN ICE SHEET

The upper limit of glacial action of the European ice sheets has been determined at several places in Sweden and Norway. The records are generally referred to the last glaciation, though it perhaps is not quite certain that they all belong there. In northern Scandinavia the last ice sheet probably overtopped the Sareks (67° 25′ N., 17° 45′ E.), whose highest peak is 2091 m. (6870 feet), although erratic boulders have not been found up to greater altitude than 1850 m. (6070 feet) (Hamberg, 1901, p. 173). On Torneträsk (68° 20′ N., 18°–20° E.) the ice covered the loftiest summit, which rises to 1800 m. (5905 feet) (Enquist, 1918, p. 7). Whether the Kebnekaise Mountains (67° 50′ N., 18° 40′ E.), which include the highest peak in Sweden of an altitude of 2123 m. (6965 feet), were overtopped by the ice sheet is uncertain (Enquist, 1918, p. 7). Thus it appears as if all the summits in the eastern part of the Scandinavian mountain range, with the possible exception of the Kebnekaise, were buried beneath the ice. As the surface of the ice fell towards the periphery, mountain peaks projected above it farther west. The most easterly

nunatak in northern Scandinavia was Rivtinn (68° 35′ N., 17° 50′ E.), 1458 m. (4783 feet) high. Since the adjacent summits present glacier action the ice surface here reached about 1400 m. (4590 feet) elevation (Enquist, 1918, p. 12). In the inner part of the Ofotfjorden (68½° N., 17½° E.) the surface of the ice sheet may have lain at 1000 to 1200 m. (3280 to 3935 feet), and at Svolvær (68° 15′ N., 14° 35′ E.) in Lofoten at 100 m. (330 feet) above the sea (Ahlmann, 1919, p. 214). In the region of Saltfjorden (67° 15′ N.) it appears to have stood about 900 m. (2950 feet) above sea level (Ahlmann, 1919, p. 214; see also Enquist, 1918, p. 13). In the tract of the Svartisen Glacier, between the 66th and 67th parallels of latitude, the continental glacier may have overtopped all peaks, some of which rise to 1600 m. (5250 feet) (Enquist, 1918, p. 14). South of Ranfjorden the Börgefjellene (65° 15′ N., 14° E.), 1660 m. (5445 feet) high, were entirely covered by the ice (Holmsen, 1913, p. 21). So were most probably also the Okstinnerne Mountains (66° N., 14° 15′ E.), which are as much as 1912 m. (6272 feet) high, although erratic boulders have been observed only to an altitude of 1800 m. (5905 feet) (Hoel, 1910, p. 24).

In western and southern Norway the altitude of the surface of the ice sheet has been determined at some places by Ahlmann (1919, pp. 140, 212–217). In the upper Romsdalen, 10 km. (6 miles) south of Stuefloten (62° 15′ N., 8° 10′ E.), it has been fixed at about 1700 m. (5775 feet), near Romsdalshorn and the Vengetinnerne in the lower part of the same valley at about 1500 m. (4920 feet), and on the southern side of the Romsdalsfjorden, opposite Molde, at about 1100 m. (3600 feet). Romsdalshorn, 1550 m. (5085 feet), and the Vengetinnerne, 1820 m. (5970 feet) high, may have formed nunataks. Mt. Gausta (59° 50′ N., 8° 45′ E.), 1880 m. (6168 feet) high, also rose above the ice; while Håteigen (60° 14′ N., 4° 45′ E.), 1680 m. (5545 feet) high and forming a monadnock in the southern part of the Hardangervidda, was buried beneath the ice (also Brøgger, 1893).

In Finland, where high mountains, that is mountains up to 1359 m. (4459 feet) occur only in the narrow wedge between Sweden

and Norway, the ice probably covered all the land. The ice certainly also buried the whole of southern Sweden, whose highest peak, Tomtabacken (57½° N., 14½° E.), is 377 m. (1237 feet), and Denmark, whose highest point, Ejer Bavnehøj (56° N., 9° 50′ E.), is only 172 m. (564 feet). On Rummelsberg in Silesia the ice surface reached an altitude of 330 m. (1085 feet) (Frech, 1912, p. 357).

## Slope of the Ice Surface in the Peripheral Belt

The foregoing data give some idea of the surface slope of the ice sheets. The inclination of the ice surface from the White Mountains to Mt. Monadnock, 95 miles (155 km.) distant, amounted to 1 in 162, or 32 feet per mile, if the mountains be supposed to have been overtopped by equally thick ice layers. The gradient from Mt. Monadnock to the terminal moraine, 120 miles (195 km.) distant, was 1 in 264, or 20 feet to a mile, if calculated from the top of the mountain to an ice surface 800 feet (245 m.) high at the moraine; or it was 1 in 198, or 27 feet per mile, if calculated to sea level at the moraine. Between the Adirondacks and the Catskills, 150 miles (240 km.) apart, the slope was 1 in 547, or 9.7 feet to a mile, if the ice is assumed to have barely overtopped Mt. Marcy. From Slide Mountain to Pokono Knob, a distance of 80 miles (130 km.), in which the ice surface fell about 1900 feet (580 m.), the gradient was 1 in 222, or 24 feet a mile.

If the figures of the upper surface of the ice are comparable to each other, the surface slope in reality was notably less, because the then land surface, as compared with the modern one, inclined from the periphery of the glaciated area towards its center.

Near Baraboo, Wisconsin, where the border of the ice during its maximum extent lay along the side of a bold ridge that records the upper limit of the ice, the average surface gradient for the last one and three-fourths miles was 1 in 16.5, or 320 feet per mile (Salisbury, 1895). In Montana the surface slope near the outermost moraines varied from 1 in 176 to 1 in 75, or from 30 to 70 feet to a mile (Calhoun, 1906, p. 28).

The average surface gradient in the lowest 70 miles (115 km.) of

the Juan de Fuca Glacier of the Cordilleran ice sheet was at least 1 in 66, or 80 feet per mile; and that in the lowest 5 miles of the adjacent Puget Sound Glacier was 1 in 112, or 47 feet to a mile (Bretz, 1920, p. 339).

From the inner part of Ofotfjorden to Svolvær, Norway, according to Ahlmann's (1919, pp. 141, 214) data, the surface of the ice sheet sank about 1000 m. in 130 km. (3280 feet in 80 miles), or with a gradient of 1 in 130. In Romsdalen the upper limit of glaciation inclined in peripheral direction from 1700 to about 1150 m. in 30 km. (5580 to 3775 feet in 19 miles), that is 1 in 55. In Jämtland, in northern Sweden, gradients of from 1 in 166 to 1 in 55 were found in the marginal belt of the waning ice sheet (Frödin, 1915). The inclination was found by Frödin not always to increase as the thickness diminished, though it generally did.

The most accurately determined surface slope in the peripheral belt of the Pleistocene ice sheets is that between Slide Mountain and Pokono Knob, which amounts to 1 in 222, or 24 feet to a mile. The minimum average gradient in New England probably was 1 in 208 to 1 in 176, or 25 to 30 feet per mile. These amounts are only one-fourth to one-half of the gradients in the marginal zone of the modern Greenlandic ice sheet (see p. 55). Understanding of these incongruent figures demands a discussion of some phases of the ice motion. This discussion, in turn, will enable us to make rough estimates of the surface slope in parts of the ice sheets from which no exact data are available.

## DISPERSAL OF BOULDERS AS EVIDENCE OF ICE FLOW FROM THE CENTRAL PARTS OF THE ICE SHEETS

Although, as held by Chamberlin and Salisbury (1906, p. 356), the growth of the Pleistocene ice sheets to a large extent took place through nourishment of an inframarginal belt, centrifugal ice flow from the very central areas is suggested by the transportation of boulders over long distances. Erratics of granitic gneiss, etc., from the Adirondacks occur in the terminal moraine in Long Island (F. J. H. Merrill in Woodworth, 1901, p. 627) and on the Pokono plateau and vicinity in Pennsylvania (G. F

Wright, 1890, p. 51; 1900, p. 211), some 225 miles (360 km.) from the source. A boulder of the very characteristic basic gabbro, or norite, that occurs in many places in the Adirondacks has been observed by the late Professor James F. Kemp west of Newburg, N. Y. Erratics of norite gneiss, etc., that may have come from some point northeast of a line from Ottawa, Canada, to the southwest side of the Adirondacks, and very probably from the area in Quebec between Ottawa and Quebec city, have been found in northern Kentucky, 650 to 700 miles (1050 to 1125 km.) from the source (Jillson, 1925). Boulders of jasper conglomerate from the region north of Lake Huron or near the lower end of Lake Superior or northern Minnesota have been found in Ohio, Kentucky, southern Indiana, Illinois, central Missouri, etc., having been transported as much as 650 miles (1045 km.) (G. F. Wright, 1890, p. 52; 1900, p. 213; Wilkerson, 1927). Native copper from the southern shore of Lake Superior has been collected in southern Illinois, some 600 miles from its source, in southern Iowa 450 miles (725 km.) distant and in northern Missouri some 500 miles distant (C. A. White in Dana, 1895, pp. 953, 959; Wilkerson, 1927). Ice-rafted copper has been found in the extreme western part of Pennsylvania (Williams, 1917, pp. v, 85). Diamonds probably from the region north of Lake Superior or possibly from the vicinity of the southern end of Hudson Bay have been found in a belt extending from Ohio across Indiana, Illinois, and Wisconsin to Minneapolis (Hobbs, 1899; Leverett, 1915, p. 65). Boulders of anorthosite from the west shore of Lake Superior have been collected in central Missouri 600 miles (965 km.) from their source (Wilkerson, 1927). Felsite erratics in North Dakota and Minnesota believed by Bell (1887, p. 36) and Upham (1896, p. 131) to come from Long Island (55° N., 79° W.) in Hudson Bay, as Dr. W. H. Collins points out, perhaps originated at some other locality in the Pre-Cambrian shield. Laurentian boulders at the foot of the Rockies in southern Alberta (49° N., 113° W.) no doubt were transported from the Pre-Cambrian area some 500 miles (800 km.) to the northeast (Dawson and McConnell, 1884, p. 149; and information by Dr. W. H. Collins).

In Europe long distance transportation of boulders by the ice sheets is well known (see Wahnschaffe and Schucht, 1921, pp. 82–92; Kummerow, 1925). It is enough here to mention that boulders of characteristic rocks in Dalarna and the Åland Islands, that is from near the center of the glaciated area, were spread in wide fans and carried to the outermost limits of glaciation in the central part of the European continent (Milthers, 1909).

The transportation need not have taken place during a single glaciation, but the first ice sheet may have carried the boulders part of the way, the second ice another part, and so on. However, the fact that, for instance, Åland erratics are spread in a fan extending from southern Norway over Germany far into Russia shows that the carriage must have proceeded very far during a single glaciation. The occurrence of erratics from the Adirondacks, etc., in the oldest drift in Pennsylvania (Williams, 1917, p. 18) and in Kentucky (Jillson, 1925) speaks eloquently for the same thing.

That there was considerable outflow of ice in the European ice sheet is also evident from the pushing together of winter moraines even during times of great summer retreat, as at the uncovering of the Stockholm region, and from uphill westward ice movement, as the ice divide during the waning of the ice was situated to the east of the Scandinavian mountain range. The ice flow prevailed till the ice sheet had nearly disappeared.

### ICE FLOW UNDER LOW GRADIENT POSSIBLE WITH HIGH TEMPERATURE OF THE ICE

Thus considerable outflow of ice from the central parts of the Pleistocene ice sheet may have taken place. During the growth and the maximum extent the forward motion of the ice in low latitudes, where depletion no doubt was great, must have been essentially faster than that of the modern ice sheets. As the surface slope of the ice sheet over New England seems to have been but one-half to one-fourth that in the marginal zone of the Greenlandic continental ice this rapid motion at first seems unexpected; but it may largely find its explanation in differences in the temperature of the ice.

Temperature measurements in the Storström Glacier (76° 40′ N., 22½° W.) in eastern Greenland show that the seasonal temperature variations decrease quickly with depth, being probably 40° C. (72° F.) at the surface, about 5° C. (9° F.) at a depth of about 6 m. (20 feet), and a fraction of a degree at that of 24 m. (79 feet), the greatest depth at which observations were made (J. P. Koch, 1916, p. 9). The practically constant temperature at that depth (24 m.) below the surface was −14° C. (6.8° F.). This is perhaps the annual mean temperature of the place. From the level of constant temperature the temperature rises with depth. The gradient appears to be 1° C. in 20 m. (1° F. in 36 feet) or possibly 1° C. in 15 m. (1° F. in 27 feet). The melting point of the ice at this place may thus be attained at a depth of between 200 and 300 m. (650 and 975 feet).

In central Greenland, at an elevation of 3000 m. (9850 feet) above the sea, the temperature of the ice a few meters below the surface was found to be about −32° C. (−25.6° F.) (J. P. Koch, 1916, p. 10). Approximate melting temperature is probably to be expected at a depth of 500 to 600 m. (1640 to 1970 feet).

Important temperature measurements were also carried out by von Drygalski (1924) in a grounded iceberg in the Antarctic from April 14, 1902, to January 19, 1903. The topmost layer to a depth of 7 cm. (2.7 inches) showed the same temperature as the air, which all the time was below freezing. The annual fluctuations decreased quickly with depth, being some 5.5° C. (9.9° F.) at a depth of 5 m. (16 feet), 0.9° C. (1.62° F.) at that of 15 m. (49 feet), and 0.7° C. (1.26° F.) at that of 30 m. (98 feet). The mean temperature at depths between 15 and 30 m. amounted to −10.1° to −10.4° C. (+13.82° to 13.28° F.) and was 1° C. (1.8° F.) higher than the mean air temperature, probably owing to the warming influence of the sea.

Thus the march of the temperature in the surface layers of glaciers and ice sheets in regions with low annual mean temperature seems to be on the whole comparable with that in the surface layer of the earth's crust, as discussed by Chamberlin and Salisbury (1905, p. 277). In the very top layer the temperature is the

same as in the air above, when this is below thawing. Beneath this thin layer the temperature fluctuations decrease downward. Their amount and the depths to which they penetrate are dependent on the difference between the winter and summer temperatures of the air, on the severity of the winters, and no doubt also on the consistency of the ice. At a depth of some meters to a few tens of meters the temperature is practically constant and probably equals or is somewhat higher than the mean air temperature, provided this is below thawing. From the zone of constant temperature the temperature increases depthward at a probable rate of 1° C. in 15 to 20 m. (1° F. in 27 to 36 feet), until the melting point is nearly reached. From here downward the ice temperature may be slightly below the melting temperature at the respective pressure, judging from results obtained from measurements in Alpine glaciers down to a depth of 153 m. (502 feet) (Hess, 1904, pp. 151–154, 320). The Alpine glaciers whose temperatures have been studied lie in regions whose annual mean temperature is at or somewhat above the melting point; and the temperature fluctuations were found to penetrate only to a depth of 8 m. (26 feet), below which the temperature is slightly below the melting point.

The physical properties of ice, of importance for ice flow, vary considerably with temperature. The hardness of Greenland ice at −15° C. (+5° F.) was between 2 and 3 and at −40° C. (−40 F.) was 4 (J. P. Koch, 1916, p. 11). The flexibility of the ice at low temperature also was found to be less than at temperatures near the melting point. The viscosity of ice diminishes as the temperature rises (see Wright and Priestley, 1922, p. 478). The nearer the temperature of the ice is to the melting point, the easier the flow, the smaller the necessary surface slope. A part of an ice body with lower temperature undergoes no motion towards a warmer part unless the gradient is considerable (Drygalski, 1897, p. 513). Thus the temperature of the ice is the chief factor in determining the necessary surface slope for flow of an ice sheet resting with full weight on the ground. As a thick ice body has a relatively greater part near the melting point, it

requires less surface slope for motion than does a thin sheet. The Storström Glacier, which according to Koch moved in spite of its low temperature, probably did so because it was a narrow tongue of the land ice.

The observed great differences in temperature between the Greenlandic ice and the glaciers in the Alps make it evident that the temperature of the upper strata of the Pleistocene ice sheets, which extended over so many degrees of latitude, must have been very different. In the marginal belt in low latitudes as in New England, where the summer temperature must have been fairly high, only a very thin surface layer may have been appreciably below the melting point and the ice consequently plastic and motion easy. In the low latitudes the ice in the marginal belts at all slopes must have been subject to thawing during the day in summer time, and part of the melted water may have frozen by night before reaching the ice edge, thus contributing to attenuate the ice. As the ice temperature fell and the flow grew more difficult in higher latitudes the surface slope must have undergone a corresponding increase. In regions like Labrador the surface inclination in the marginal belt may have been comparable to that in Greenland.

The conditions touched upon may largely explain how the ice flow in New England could be faster than in Greenland in spite of a much less inclination of the ice surface. We may therefore assume that the observed minimum slope of the ice was the actual one, that from the terminal moraine to the White Mountains, lying some 200 miles inside, it was 1 in 176, or 30 feet to the mile, and that the altitude of the ice at the White Mountains was about 6500 feet.

## Slope and Altitude of the Ice Surface in the Interior

The Cordilleran ice cap along the 49th parallel, where it was 250 miles (400 km.) wide, had a practically flat surface (Daly, 1912, p. 577, Pl. 49). In British Columbia, also, according to Mr. W. A. Johnston, it was remarkably even.

In the interior of the great North American ice sheet the sur-

face slope perhaps equaled that in the interior of Greenland today. The larger extent and the greater thickness, to be sure, suggest less inclination; but, on the other hand, the more rapid centrifugal ice motion suggests greater slope and considerable outward pressure. Inside the White Mountains the slope therefore probably averaged about 10 feet to a mile, or 1 in 528. In the 800 miles (1290 km.) from the mountains to the Labrador ice center the surface then may have risen some 8000 feet (2450 m.), and at the center reached some 15,000 feet (4600 m.). The correctness of this figure is suggested by the reasonable gradient of the ice surface it gives towards the northeast. The average gradient to the Nakvak Fiord would be 1 in 115, while the mean gradient of the Greenlandic ice to its highest point along J. P. Koch's route is 1 in 133.

In Montana the ice surface reached an altitude of more than 4500 feet (1370 m.) (cf. p. 62). The distance to the Keewatin ice center is the same as that from Long Island to the Labradorean center, or 1000 miles (1610 km.); and if the same gradients be assumed a height of the ice at the Keewatin center of some 18,000 feet (5500 m.) is obtained.

Since the central areas were considerably pressed down, while the peripheral region took about the same level as today, the figures may at the same time approximately give the actual thicknesses.

The altitude of the central part of the European ice sheet can be approximately estimated in two ways. Suppose the surface slope towards the southeast was the same as it was southward in the Labradorean ice. The surface 200 miles (320 km.) inside the border then reached 6000 feet (1830 m.) above the base of the edge; and 500 miles (800 km.) still farther in, or in the center, it reached 11,000 feet (3350 m.) above the base of the margin. The land at the periphery being low and the central area depressed, the thickness at the center may have amounted to some 11,000 feet (3350 m.).

During part of the glaciation, probably during its climax, the ice shed in northern Sweden was situated to the east of the

mountain range. The position is not well known but was probably about halfway between the Norwegian frontier and the Gulf of Bothnia (Frödin, 1925, pp. 185, 209, Pl. 6). It is not impossible that it even lay over the Gulf of Bothnia (Enquist, 1918, pp. 27, 82, Fig. 34). From the ice shed the ice moved towards the west and northwest uphill and across the Scandinavian mountain range. This fact implies that the slope of the ice surface was greater than the opposite inclination of the ground (see Brückner, 1914) and thus enables a rough estimate of the elevation of the ice in the center. Since the average altitude of the land on the watershed is about 1000 m. (3280 feet), since the surface of the ice here reached an altitude of about 2000 m. (6560 feet), and since the ice shed lay on land some 300 to 400 m. (1000 to 1300 feet) high, the central parts of the ice sheet must have reached an elevation of at least 2700 m. (8850 feet) plus an amount corresponding to the surface slope necessary for ice motion on a level substratum. If this gradient be assumed to be as low as 1 in 352, or 15 feet to a mile, some 1350 feet (400 m.) have to be added for the 90 miles (150 km.) the ice shed lay east of the watershed. As, furthermore, the central parts were suppressed at least one or two hundred meters more than the mountain chain, the ice on the ice shed must have reached an altitude of nearly 3500 m. (11,480 feet) and have been over 3000 m. (9840 feet) thick.

It should be pointed out that, since a much lower gradient is assumed for the marginal belt than that prevailing in Greenland at present, the calculated altitudes of the ice surface are relatively much less than those observed in Greenland. Thus the European ice sheet with the same relation between width and height as the Greenlandic ice cap would have reached an altitude of 5000 meters.

These estimated gradients and heights of the surfaces of the Pleistocene ice sheets are perhaps nearly correct. They are scarcely too large, but possibly too small, too conservative, for it seems somewhat doubtful whether they can explain the strong ice motion that appears to have taken place. The ice flow in the

margin in low latitudes in spite of the small surface slope perhaps required great pressure from the interior and thus greater inclination of the ice surface there than now exists in the modern ice sheets. These are survivals from the glacial epoch. In the Greenlandic ice, supply and wastage barely balance, and the Antarctic ice is probably shrinking (see pp. 56, 59). Their inability to expand is perhaps chiefly due to insufficient nourishment of the interior parts, which now are too low.

# CHAPTER IV

## THICKNESS, EXTENT, AND VOLUME OF THE PLEISTOCENE ICE SHEETS AND GLACIERS

Computations of the volume of ice that there was in mountain glaciers and ice sheets at the climax of the last glaciation, must of necessity be even less exact than estimates of thickness based upon the figures in the preceding chapter. Not only are the areas of glaciers imperfectly known and measured, especially outside of North America and Europe, but as we now come to multiply two uncertain dimensions by a third we shall multiply the errors of all three. Nevertheless, in this chapter we shall take all the facts we have and work out the volumes as well as we can.

### North American Ice Sheet

In Chapter III figures are given for the actual or the minimum height of the ice sheet at different points during the stage when it was thickest. The actual thickness, of course, was as much less as the surface of the underlying land stands above sea level. The approximate average thickness near the ice edge may have been somewhat less than 1000 feet (300 m.); and the average minimum thickness may have been 1500 to 2000 feet (460 to 610 m.) in southern and central Massachusetts, 2000 to 2500 feet (610 to 760 m.) in southern New Hampshire, 5000 feet (1525 m.) in northern New Hampshire and Vermont and the interior of Maine, and more than 1500 feet (450 m.) at Mt. Desert Island on the coast of Maine. Since, however, the ice probably overtopped all the mountains in New England, all these thicknesses perhaps were greater.

In easternmost Pennsylvania, northern New Jersey, and southern New York, at some distance from the terminal moraine, the approximate mean thickness may have been 1000 feet

(300 m.); in the northeastern corner of Pennsylvania it may have been 1500 feet (460 m.), on the latitude of the Catskills about 2500 feet (760 m.), and in northern New York perhaps some 5000 feet (1525 m.).

In the Magdalen Islands the thickness of the ice probably was about 200 feet (60 m.), in the southeastern part of Gaspé Peninsula 1700 feet (520 m.), and in the southwestern portion of the same peninsula 2500 feet (760 m.) (Coleman, 1920, pp. 322, 324).

The Labrador ice sheet where it reached the sea, between the Strait of Belle Isle and Saglek Bay, was relatively thin, certainly not more than 2000 feet (610 m.) at Mugford and Hebron for instance (Coleman, 1921, p. 28). In the vicinity of the Torngat Mountains the ice was over 2650 feet (810 m.) thick (Daly, 1902, p. 251).

On the southern shore of Lake Superior the ice was at least 1250 feet (380 m.) thick measured from the present water surface of Lake Superior, for the Huron Mountains were overtopped (F. B. Taylor, 1915, p. 509). In northern Montana, at the West Butte of the Sweet Grass Hills, 25 miles (40 km.) from the outermost edge of the ice, its thickness was about 1300 feet (400 m.), the hills mentioned forming nunataks (Calhoun, 1906, p. 28). In the region of the Cypress Hills, Saskatchewan, 250 miles (400 km.) inside the outer edge, the average thickness was 1500 feet (460 m.), the maximum thickness 2000 feet (610 m.) (McConnell, 1886, p. 76).

In the Keewatin ice center the inferred thickness was some 18,000 feet (5500 m.) and in the Labradorean center about 14,000 feet (2270 m.).

According to the data at hand the approximate mean thickness of the ice along a line from Long Island to the Labrador ice center was 8200 feet (2500 m.), along a radius from Montana to the Keewatin center 10,650 feet (3250 m.), and along a line from Nakvak Fiord, northeastern Labrador, to Labrador center 9000 feet (2750 m.). The mean ice thickness in the area east of the Rockies thus may have been nearly 9000 feet (say 2700 m.).

In the western mountainous regions the thickness, of course,

was much less. The average thickness of the Cordilleran ice sheet on the 49th parallel of latitude was 2500 feet (760 m.) (Daly, 1910, p. 299). Farther north the thickness must have been greater, for the ice flowed south and southwest (W. A. Johnston). The mean thickness of the ice sheet to the west of the east side of the Rockies may have been over 3500 feet (say 1100 m.).

The extent of the last ice sheet in the north is not well known. According to Chamberlin's (1913) map the whole Arctic archipelago was glaciated during the Pleistocene, though not necessarily during the last glacial epoch. According to Schuchert's (1914, p. 266) and Alden's (1924, p. 32) maps, on the other hand, fairly large parts of it escaped glaciation. The latest explorations seem to show that northern Baffin Land, Melville Peninsula, and Southampton Island were entirely buried beneath ice, as held by Chamberlin (Freuchen and Mathiassen, 1925, pp. 552, 555, 560). As the evidence includes striae, it seems probable that the glaciation was synchronous with the last ice sheet.

Thus the area of the last North American ice sheet may have comprised nearly 4,500,000 square miles (say 11,500,000 square km.). The area to the east of the Rockies may have been some 3,500,000 square miles (say 9,000,000 square km.), and that to the west of the eastern front of the mountains less than 1,000,000 square miles (say 2,500,000 square km.). With the mean thickness of the ice in the former area 2700 m., the volume of the ice here amounted to 24,300,000 cubic km. Assuming the ice in the latter area to have been some 1100 m. thick on an average, its volume was 2,750,000 cubic km. The whole volume of the last North American ice sheet thus estimated was about 27,050,000 cubic km. (6,489,000 cubic miles).

### European Ice Sheet

The central parts of the glaciated area in Europe, or the region somewhat to the west of the Gulf of Bothnia, with an average altitude of a few hundred meters and the ice surface standing at an altitude of nearly 3500 m., had an ice cover which over a large

area may have been between 3000 and 3500 m. (9840 and 11,480 feet) thick. On the west side of the Scandinavian mountain range, where the ice surface sank from a 2000 m. elevation to sea level and the average altitude of the ice was 1000 m. and the mean height of land more than 500 m., the average thickness of the ice probably was 500 m. (1640 feet) or somewhat less. South of the Trondhjemsfjorden, Norway, where the highest elevation of the ice was 1700 m. and the height at Molde 1100 m., and where the mean altitude of the land is 500 to 600 m., the average thickness of the ice may have amounted to 700 or 800 m. (2300 or 2625 feet).

In Finland south of the 65th parallel, which has an average elevation of 100 to 125 m., the ice thickness probably amounted to 2500 to 3000 m. (8200 to 9840 feet); and in southern Sweden, which has a mean height of about 200 m., it may have amounted to 2000 to 2500 m. (6560 to 8200 feet). At the terminal moraine in Silesia and Poland the thickness of the ice probably was about 200 m. (660 feet), even low mountains forming nunataks (Lozinski, 1909; Frech, 1912, pp. 352, 357).

The average thickness of the ice sheet during its greatest extent along a line from the center southeastward into Russia may have been approximately 2000 m. (6560 feet). This may also have been the mean thickness of the whole part of the ice situated to the southeast of the Scandinavian mountain range. In the highlands of western Sweden and in Norway the mean thickness probably amounted to 500 to 600 m. (1640 to 1970 feet).

The last European ice sheet covered an area of approximately 3,300,000 square km. (1,274,130 square miles) (A. Penck, 1922, p. 308). The part to the east of the mountain range represented about two-thirds, or 2,200,000 square km. The volume of this eastern part consequently amounted to 4,400,000 cubic km., that of the western part to 550,000 to 660,000 cubic km., and the whole ice sheet to some 5,000,000 cubic km. (1,199,000 cubic miles).

## Other Pleistocene Glaciers in Eurasia

Besides the northern European ice sheet there were in Europe and Asia large ice caps and clusters of glaciers. The arctic lands of Spitsbergen, King Charles Land, Franz Josef Land, Iceland, etc. were much more heavily glaciated than at present, and the Scottish Highlands, the Pyrenees, the Alps, the Caucasus, the Himalayas, the ranges diagonally crossing Asia, and other high mountains carried extensive glaciers. The combined area of these glaciers may have been larger than that of the last European ice sheet. The mean thickness of the ice that in postglacial time has disappeared from these glaciers may not be great, though at places it amounts to several hundred meters. If it is 100 m., the volume of the ice probably amounts to some 350,000 cubic km.

## Pleistocene Greenlandic Ice Sheet

During the Pleistocene ice age the Greenlandic ice sheet had much greater extent and at least near the present margins was much thicker than now. The mean thickness is now probably some 1400 m. (4600 feet) (Meinardus, 1926, p. 99). Aside from small areas and solitary nunataks it probably covered the whole continent. No distinction has yet been made between the records of the several glacial epochs.

In the Upernivik District (73° N., 56° W.), on the middle of the west coast, solitary high peaks may have towered above the Pleistocene ice surface (*Meddelelser om Grønland*, Vol. 60, p. 441). In the region of latitude 71° N. and longitude 52° W. striae and erratic boulders have been traced up to elevations of between 1200 and 1600 m. (3940 and 5250 feet), above which the highest peaks probably rose as nunataks (*loc. cit.*, p. 350). At latitude 67° N. and longitude 53° W. the ice covered all lands, and its surface stood at an altitude of 600 to 800 m. (1970 to 2625 feet) above the sea (*Meddelelser om Grønland*, Vol. 61, p. 16). At latitude 67° N. and longitude 49° 40′ W. a nunatak situated 6 km. (4 miles) inside the edge of the ice sheet and rising 375 m. (1230 feet) above it has till on the top (*loc. cit.*, p. 109). In the

Godthaab District (64° N., 52° W.), in southwestern Greenland, ice covered all land up to altitudes of 950 to 1250 m. (3120 to 4100 feet), so that only few points formed nunataks (Kornerup, 1890, p. 105; *Meddelelser om Grønland*, Vol. 61, p. 193). The land from Merkuitsok (63° 45′ N.) down to Frederikshaab Isblink (62° 30′ N.) was covered by ice whose surface at the inner part of the fiords stood at elevations of 940 to 1100 m. (3080 to 3610 feet) and near the coast at altitudes of 250 to 660 m. (820 to 2165 feet) (Kornerup, 1890, p. 112). A few peaks rose above the ice. In Frederikshaab District (62° N., 50° W.) the ice surface stood at least at an altitude of 500 m. (1640 feet) and in the southernmost part of the continent at that of 1200 m. (3940 feet) (*Meddelelser om Grønland*, Vol. 61, pp. 313, 426). The northernmost part of Greenland was ice-free (Lauge Koch, 1925, pp. 272, 283).

Inside a narrow coastal plain the land rapidly rises to a strongly broken highland with peaks 500 to 1200 m. (1640 to 3940 feet) high, so that the average thickness of the Pleistocene ice in the now ice-free regions may have been 400 m. (1310 feet) or somewhat more. As this area is about 300,000 square km. this represents some 120,000 cubic km. of ice. In the interior the thickness must have been greater than now, though, of course, no record of this exists. On an average we may suppose it to have been 100 or 200 m., say 150 m., greater. Taking the area of the existing ice sheet as 1,869,000 square km. (Hammer, 1921, p. 1), the volume of the lost ice may be figured as 280,000 cubic km. The quantity of ice that has disappeared from the whole continent since the last climax then probably amounts to some 400,000 cubic km.

### Pleistocene Glaciers in the Southern Hemisphere

According to David (1914, p. 622) the Antarctic ice sheet, since its last climax, which may have been synchronous with the last Pleistocene glaciation in the southern hemisphere (see p. 13), has probably lost an ice layer some 300 m. (1000 feet) thick. As the area is about 13,500,000 square km. (over 5,000,000 square

miles) this represents some 4,050,000 cubic km. of ice. Since, too, the ice sheet certainly had then a greater extent than now, the lost quantity is probably still greater. The average modern thickness of the ice sheet according to David (1914, p. 613) is probably about 5000 feet (1525 m.) and according to Meinardus (1925, p. 188) at least 1000 m. (3280 feet).

The extent of the last glaciation in South America is not known, but let us assume that the amount of lost ice in the southern hemisphere outside the Antarctic is 50,000 cubic km.

# CHAPTER V

## STAND OF SEA LEVEL AT THE CLIMAX OF THE LAST GLACIATION

The volume of ice during the climax of the last glaciation in *excess* of the existing quantity, according to the estimates made in the foregoing chapter was as follows:

| | Cubic kilometers of ice |
|---|---|
| North American ice sheet | 27,050,000 |
| European ice sheet | 5,000,000 |
| Other Pleistocene glaciers in Eurasia | 350,000 |
| Greenlandic ice sheet | 400,000 |
| Northern hemisphere | 32,800,000 |

This total ice volume corresponds to 30,077,000 cubic km. of water. Taking the area of the oceans as 361,100,000 square km., this water quantity represents a layer over that area 83 m. (272 feet) thick.

The ice sheets and glaciers on the northern hemisphere are thought to have reached their greatest extent at practically the same time (see p. 8). On the other hand the climaxes of the glaciations were perhaps not entirely synchronous on the different hemispheres, though alternation is out of question (see p. 13). The volume of the ice on the southern hemisphere in *excess* of the present quantity is estimated to have been some 4,100,000 cubic km., which corresponds to 3,760,000 cubic km. of water and represents a layer over the area of the oceans 10 m. (33 feet) thick. Therefore, if the glaciations reached their climax simultaneously on both sides of the equator the sea level was lowered by some 93 m. (305 feet). If the contemporaneity was only partial the sea level may at most have been lowered 88 m. (290 feet).

The results obtained through this elaborate study accordingly agree fairly well with the moderate figures given by Daly, W. B. Wright, and W. J. Humphreys. The earlier estimates and guesses, ranging from less than 1 mm. to 914 m., have recently been reviewed by Daly (1925, p. 285).

The amount of the lowering of sea level due to glaciation thus computed may be checked by studies of the submarine topography in regions that have undergone no movement or but slight vertical movements, in late-Quaternary times. The relatively recent channel of the Hudson River on the continental shelf extends out to the depth of some 43 fathoms (258 feet; 79 m.) (cf. p. 86). It has, of course, been filled out somewhat since its formation. In the Adriatic Sea river valleys extend down to 110 m. (360 feet) below sea level (De Marchi, 1922). The topography of the shallow Sunda Sea suggests that it has partly been carved out at epochs when the sea level stood at least 240 feet (73 m.) lower than at present (Molengraaff, 1921, pp. 100, 116). These conditions may have bearing on the Pleistocene position of the ocean level.

The accurate fixing of the lowest stand of sea level is of great importance not only for the solution of problems in the fields of geology, physiography, and plant and animal geography, but also for the study of the remains of early man who frequently preferred to live near the seashore, that is to say, below the modern sea level during the glacial epochs (see Daneš, 1925; Ramsay, 1925). No less important but even less possible of full and accurate solution, until we accumulate more facts like those in Chapter VII, is the problem of the rate of rise of the oceans as the ice melted (Antevs, 1928).

# CHAPTER VI

## CONDITIONS DURING THE CLIMAX AND RETREAT OF THE LAST ICE SHEET IN EASTERN NORTH AMERICA

Studies of glacial sediments, moraines, raised beaches, and other physiographic features in areas visited between 1921 and 1926 and the consideration of consequences of the rise of sea level discussed in the preceding chapter point to conclusions regarding late glacial geography in New England and adjacent parts of Canada which it is the purpose now to state.

The chapter does not pretend to reconstruct conditions in all parts of this region, nor to treat all the conditions with equal detail; but it aims to present a new view of old problems.

### Southern New England and Adjacent Regions

During the greatest extent of the last ice sheets and the early days of waning, the sea level stood considerably lower than at present, because large quantities of water were locked up in ice. At the climax it probably was some 300 feet (90 m.) lower (see p. 81). Provided the marine limit was recorded as the uncovering took place, which is likely, the zero isobase of the late-Quaternary changes of level which runs south of Boston (Goldthwait, 1922) marks points that at the release from the ice stood exactly as much below their modern level as the then sea level stood below the present one. Outside the zero isobase the late-glacial marine shore line, the southward continuation of the raised highest marine limit, is now drowned. In other words, the belt outside the zero isobase stood higher in relation to the sea level of that time than it now stands in relation to the present sea level. The amount increased outward. If the slant of the submerged part of the late-glacial shore line is graphically determined from the raised portion of it, and if the stand of the sea level at the climax of the

glaciation was 300 feet (90 m.) below the modern, the warped old shore line would reach its horizontality, its hinge line, some 90 miles (145 km.) south-southeast of Nantucket, at latitude 40° and longitude 69½°. The hinge line probably ran from this point west-southwestward through the point 39° N. and 73½° W. and east-northeastward just outside the edge of the continental shelf off Sable Island. This tract then at the greatest extent of the ice stood in the same vertical position as it does today. The isocatabases of 100 and 200 feet (30 and 60 m.) may have run through the elbow of Cape Cod peninsula and the point 41° N. and 69½° W.; in other words, the postglacial submergence at these places may have amounted to 100 and 200 feet.

On Narragansett Bay, Rhode Island, at localities 141 and 142 varved glacial clays undoubtedly deposited in perfectly fresh water occur below and at sea level, and at locality 143, Providence, at 30 to 40 feet (9 to 12 m.) above the sea. The localities lie in open exposures to the sea. The absence of marine fossils in the late-glacial deposits of Narragansett Bay as well as of Long Island Sound was pointed out by Dana in 1875 (pp. 432, 280). The deformed shore line of the water body in which the clays and the sand and gravel deposits of Narragansett Bay were laid down was studied in 1907 by Robert W. Sayles, who has kindly put his data at the writer's disposal. The shore line at Providence lies at an altitude of about 120 feet (37 m.). It sinks uniformly southward and in latitude 41½° N., or near the mouth of the bay, reaches to modern sea level. This southward tilt and the submarine topography make it probable that the barrier of the lake in which the sediments were formed was situated on the line Montauk Point-Block Island-Marthas Vineyard. An isthmus here shows that the region stood at least 120 feet higher in relation to sea level, which is reasonable. The iceward (northward) inclination of the earth's crust at that time and the relatively lower areas inside the line mentioned may have caused the formation of a lake. As the ice border receded the lake extended up Narragansett Bay, Long Island Sound, and the river valleys to the north. This fresh-water body also accounts for the distinct

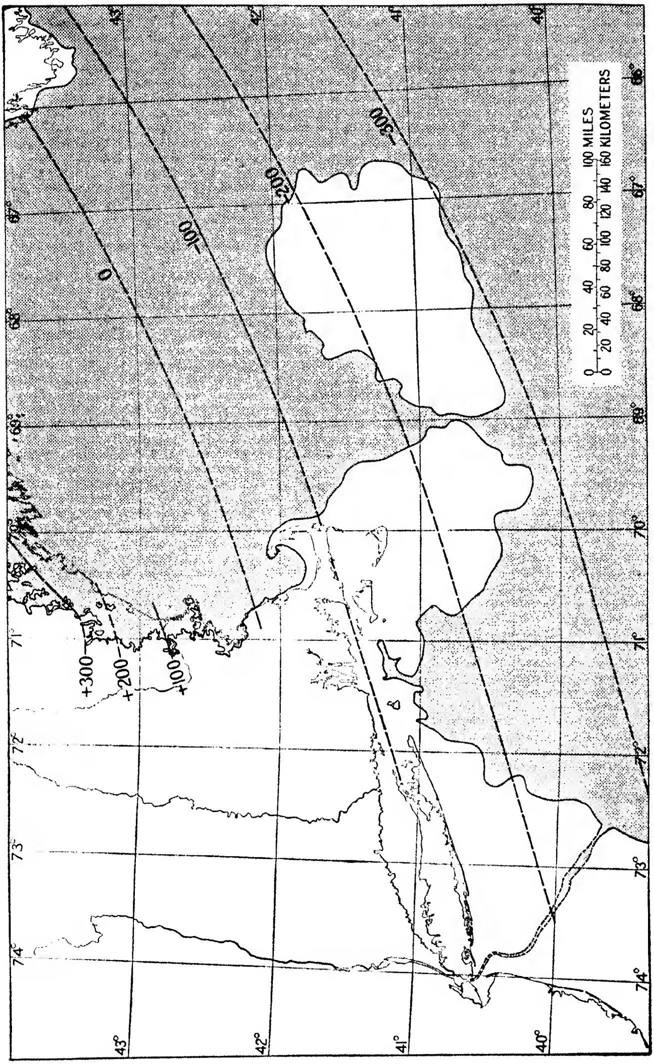


FIG. 1.—The probable glacial coast line in New England and adjacent areas and isobases and isocatabases for every 100 feet.

varvity of the late-glacial clays in southern Connecticut, as at New Haven. Lakes persisted in the river valleys at least until after the uncovering of northern New England. (Fig. 5, p. 108; Fig. 12, p. 118).

South of Hackensack, New Jersey, late-glacial fresh-water clays occur from slightly above sea level to at least 100 feet (30 m.) below it (Salisbury, 1908, p. 18). At Oradell, five miles north of Hackensack they reach the altitude of about 20 feet above sea level, and still farther north, at the state boundary, that of 30 feet (9 m.). The lake level evidently reached higher. Salisbury (1908, pp. 17, 18, 20; also Reeds, 1927, p. 60) mentions probable deltas at 100 feet (30 m.) altitude southwest of Newark, and at 80 to 100 feet (25 m. to 30 m.) altitude at and north of Hackensack. He also describes very fine gravel and sand, which must have been transported by waves, from the region north of the Raritan River (at 40° 35′ N., 74° 25′ W.), outside the terminal moraine. Woodworth (1905, p. 91) observed a glacial delta at 30 feet (9 m.) elevation in the vicinity of Perth Amboy. (Fig. 5, p. 108; Fig. 6, p. 110).

The barrier of the late-glacial lake in the Hackensack region probably lay on the line Sandy Hook-Coney Island-Long Island, which tract needs only to have stood some 30 feet (9 m.) higher in relation to sea level for this purpose. However, the southern lake shore possibly lay farther out, though not far outside the 150-foot (46 m.) curve (cf. Antevs, 1922, p. 7).

On the submerged coastal plain, as has long been recognized, there is a relatively young channel of the Hudson River extending out to 39° 40′ N. and 72° 40′ W., or 95 miles (153 km.) outside Sandy Hook. It has a maximum depth of 43 fathoms (258 feet; 79 m.) and causes a long river-like reëntrant of the 40-fathom curve. As the shore line at the mouth of this channel according to the views here advanced would lie at about 250 feet, the channel may have been cut by the Hudson during the last glaciation. Thus there probably was a long outlet river for Lake Hackensack. This would account well for the fact that the barrier lasted for several thousand years.

The deformation of the earth through the weight of the Pleistocene ice sheets, as Daly (1925, p. 311; 1926, pp. 191–200) has emphasized, was both elastic and isostatic. Both kinds of distortion should consist in depression under the ice sheets and slight bulging at a distance from them. The isostatic adjustment, according to Daly (1925, p. 308) may have taken place largely through undertow at great depth. The principal movement at the surface and within a thick shell below it, according to this view, was radial shearing, and the dominant displacement at depth (of some 1800 km.) was horizontal flow, adjusting the isostatic balance. The depression due to isostatic flow and the elastic depression may have been of about the same magnitude (*loc. cit.*, p. 311). During the greatest extent of the ice sheets the belt just outside the ice margin was not elastically depressed but was very close to its preglacial position (*loc. cit.*, p. 309), which as just shown for the region south of New England was actually the case. "With the unloading, due to the melting of the ice, the glaciated tract and a broad belt of ground surrounding it are immediately uplifted by the earth's elastic response. Both the central basin and the marginal belt of high ground persist until the rigidity of the elastico-viscous earth has sufficiently decayed. On this hypothesis, then, the marginal belt of superelevated ground dates essentially from late-Glacial time to the later part of post-Glacial time and was a product of the earth's elastic response to unloading; although a small fraction of the superelevation may have been caused by viscous flow near the surface. . . . The ultimate restoration of isostatic balance by a reversal of the 'punching' process was a natural next step, according to the newer hypothesis. The basined part of the crust was sheared up and the marginal belt of high ground gradually 'snapped' back to its pre-Glacial position" (Daly, 1925, p. 309). Thus it is theoretically probable that a belt outside the glaciated tract during late-glacial and early postglacial time took a somewhat higher absolute vertical stand than at present. A possible evidence of a bulge is the Susquehanna "deeps," described by E. B. Mathews, the formation of which seems to

demand greater velocity and then steeper gradient of the river (Daly, 1920, p. 314).

Therefore, as the sea level stood about 300 feet (90 m.) lower than now, as the continental shelf some 90 to 110 miles (145 to 177 km.) outside the terminal moraine stood where it does today or at a higher level, and as the earth's surface from here inclined gradually towards the center of glaciation, large parts of the now submerged coastal plain then formed land. A reconstruction of the conditions there would be as follows: New Jersey probably extended some 80 miles (130 km.) beyond the modern coast line, and the western part of Long Island also reached far out.[1] South of Rhode Island there was a wide bay of the sea to the vicinity of Block Island. Massachusetts stretched out in a wide lobe

[1] In the coastal regions of southern New Jersey, Delaware, Maryland, etc. the climax of the last glaciation may have been represented by the post-Talbot-Cape May erosion interval, during which the land stood much higher in relation to sea level than in the Talbot-Cape May age and at present, and the seashore had retreated far beyond the present shore line, so that streams were able to cut moderately deep valleys in the Talbot terrace (Shattuck, 1906, pp. 133, 134, 137; Pl. 31).

The view that the bulk of the Talbot-Cape May formation corresponds to the climax of the last glaciation as held by Shattuck (1906, pp. 96, 125, 127) and by Salisbury and Knapp (1917, p. 162) seems to have a weak basis. Whatever the origin of the formation, which reaches an altitude of 40 to 50 feet (12 to 15 m.) above the sea (Shattuck, 1906, pp. 66, 127; Salisbury and Knapp, 1917, p. 162), whether it is marine (as held by W J McGee, Shattuck, and others) or is partly of subaerial and partly of marine or estuarine origin (as held by Salisbury and Knapp), the region during its deposition must have stood at least 30 to 50 feet lower relative to sea level than now (Salisbury and Knapp, 1917, pp. 3, 162). Therefore, and because the seashore at the maximum extent of the ice probably stood some 80 miles (130 km). farther out than now, the beds cannot well date from this stage.

The numerous plant and animal remains found in the Talbot beds of Maryland—all the animal remains described in the monograph on the Pleistocene of Maryland come from the Talbot (W. B. Clark, 1906, p. 139)—and the few marine shells from the Cape May deposits of New Jersey also bear witness decidedly against contemporaneity of the formation with the climax of the glaciation. Vascular plants are represented by the following specifically determined forms (Hollick, 1906): *Pinus echinata*, *P. strobus*, *Taxodium distichum*, *Alnus rugosa*, *Fagus americana*, *Robinia pseudacacia*, *Nyssa biflora*, and *Vaccinium corymbosum*. *Pinus strobus* at present occurs among the Alleghanies down to Georgia, and all the others are distributed down to Florida. On the other hand the modern northern limit of distribution of *Taxodium distichum*, *Robinia pseudacacia*, and *Pinus echinata* is in Delaware (possibly in southern New Jersey), southern Pennsylvania, and southern New York, respectively. "Nearly all the marine [animal] forms [represented in the Talbot] are found living at the present time off the coast of Maryland and Virginia. A striking exception is to be found in *Rangia cuneata* (Gray) which is found only at the present time in the Gulf of Mexico from Alabama to Vera Cruz, Mexico." "*Leda acuta* (Conrad) is likewise a characteristic southern species, being common along the

southeastward to some 80 miles (130 km.) south of Nantucket. To the east of here, probably separated by a narrow strait, was an extensive land area comprising approximately the parts of

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southern coast of the United States. On the other hand, *Macoma calcaria* (Gmelin) and *Astigena elevata* (Stimpson) are distinctly northern forms, the former ranging from the Arctic regions southward to Long Island Sound and the latter being chiefly confined to the coast between Cape Cod and New Jersey" (W. B. Clark, 1906, pp. 146–148). Mammals are represented by teeth of mastodon (*Mammut Americanum*), northern mammoth (*Elephas primigenius*), and southern mammoth (*Elephas colombi*) (Lucas, 1906). All three are represented in the Quaternary from New England or southern Canada down to Florida (Hay, 1923, pp. 415, 429, 431). In the corresponding beds in southern New Jersey marine shells have been found at a few places up to 20 feet (6 m.) above sea level (G. H. Cook, 1857, pp. 26–29; Merrill, 1885, p. 71; Chamberlin and Salisbury, 1906, p. 451). The species found are *Venus mercenaria*, *Ostrea virginica*, *Urosalpinx cinereus*, and *Ilyanassa obsoleta*. The first three range in modern time from the Gulf of St. Lawrence, Prince Edward Island, and Nova Scotia, respectively, down to Florida. *Ilyanassa* is common in the Quaternary from Massachusetts to South Carolina (W. B Clark, 1906, p. 183).

Thus both flora and fauna indicate clearly that the climate during the deposition of the Talbot-Cape May beds was at least as genial as today. No doubt, therefore, the formation dates from an interglacial epoch and then almost surely from the last one, though it is believed by Hay (1924a, p. 259, also 1927) to belong to the first interglacial, as it contains, all along the coast from Staten Island to Mexico, what he regards as a first interglacial vertebrate fauna.

The two other Quaternary formations of the Coastal Plain, separated from the Talbot-Cape May and from each other and preceded by epochs of erosion, the Wicomico-Pensauken and the Sunderland-Bridgeton which is the oldest (Shattuck, 1906; Salisbury and Knapp, 1917), probably date from two earlier interglacials, while the three erosion intervals may indicate three glacial epochs preceding the last one, the Wisconsin. A Wicomico deposit in the city of Washington also contains organic remains indicating somewhat milder climate than the present and is attributed to an interglacial stage (Wentworth, Berry, Mann, LaForge, 1924, pp 10, 14, 15, 40). While, consequently, the alternations of deposition and erosion may at least largely have been due to oscillations of sea level as the Quaternary ice sheets waxed and waned, the land must have been subjected to a very slow uplift, since the Wicomico and the Sunderland formations in Maryland reach altitudes of about 100 and 220 feet (30 and 67 m.) respectively (Shattuck, 1906, pp. 73, 71).

The transgression leading to the modern state of affairs (Shattuck, 1906, pp. 66, 134, 137) no doubt was essentially brought about by the rise of the sea level, as the last ice sheets melted and the water was restored to the ocean.

The views here advocated are in the best of agreement with Sayles's and T. H. Clark's (Sayles, 1924; Sayles and Clark, 1925) find in Bermuda of three distinct fossil soils, 2 to 10 inches (5 to 25 cm.) thick, intercalated between strata of eolian limestone, 30 to 50 feet (9 to 15 m.) thick, all of Quaternary age. The eolian beds are by Sayles considered to have been deposited during epochs of high stand of the land relative to sea level, when Bermuda extended over a large shallow-water area of loose material to the west of the island. The soils are believed to have been formed during periods when the climate and the relation of land to sea were about as today. Thus, Sayles holds, no doubt correctly, that the oscillations of the shore line were largely due to changes of sea level as the sheets grew and melted, and he correlates the eolian limestone beds with the glacial epochs and the soils with interglacial episodes. The three fossil soils agree well with the three Quaternary formations of the Coastal Plain.

Georges Bank that lie between the 45-fathom curve on the southern side of the bank and the 25-fathom curve on the northern. Nova Scotia extended farther south. The Sable Island Bank formed a large island, and so did the bank to the northeast of this. The remarkably even Great Bank of Newfoundland, which now reaches to within 40 fathoms (73 m.) of the sea level on an average, may have formed land for a short time. Whether the other banks south of Newfoundland rose above sea level is still more uncertain.

This belt on and near the outer hinge line is supposed to have undergone little or no change of level. Part of it probably rose a little when the ice had retreated some distance and in post-glacial time sank again, but it has chiefly become submerged by the rise of the sea level as the ice sheets and glaciers melted. This was a very slow process going on till a few thousand years ago.

Corroborating evidence that part of the drowned coastal plain in late-Quaternary time formed land is furnished by the flora of eastern Canada and Newfoundland. This contains a number of southern forms, cut off from their continuous range of distribution, whose presence can only be explained by the assumption of a now submerged route of immigration from New Jersey (Fernald, 1911, 1915, 1921). In southern Newfoundland (south of the North Peninsula) there is "a large number of species of southern origin, some even of tropical affinity. Further, on the shores of Dawson's warm Acadian Bay, including the region from Cape Breton to the south side of Baie des Chaleurs, i. e. eastern Nova Scotia, eastern New Brunswick, Prince Edward Island, and the Magdalen Islands, we find many such species, some identical with, others different from those reaching Newfoundland, and on Sable Island there are some similar cases, some of the species from the south getting to Sable Island but not to the other areas" (Fernald, 1915). In Nova Scotia there are a great number of such isolated outliers of southern and tropical plants whose northern limit of distribution otherwise lies in New Jersey, Massachusetts, southern New Hampshire, etc., and which are

entirely unknown in the regions between, or represented only at a few intervening localities like Cape Cod. Among these may be mentioned *Ilex glabra*, *Schizaea pusilla*, *Lophiola*, *Utricularia subulata*, *Eleocharis tuberculosa*, *Polygonum robustius* (Fernald, 1921, Pl. 130, pp. 274, 186, 243, 291, 233, 260). Many of these forms having decidedly southerly distribution, it is inconceivable that they should have become dispersed to these remote areas under conditions much colder than at present and rather probable that they reached them during a period of somewhat higher temperature than today (Fernald, 1915). Accordingly part of the now submerged coastal plain may still have been land when the temperature had risen higher than that now prevailing.

The frequent occurrence of marine shells in the lower strata of the late-glacial clays in Maine and the great rarity of them in the upper layers of the same clays, as Stone (1899, pp. 55–58) points out, may indicate decreasing salinity of the water. This can hardly be explained by the melt water, since this must have primarily affected the bottom layers. The condition was perhaps chiefly due to rise of the region Nova Scotia-Georges Bank and the narrowing of the inlet to the extensive and deep bay inside.

Many plants confined, at least in northeastern North America, to estuaries are believed to have required brackish or nearly fresh water for their dispersal and to have become unable to spread further because the water had grown too saline (Fassett, 1925, p. 81). The remote time of this isolation is "indicated by certain species and varieties which are confined to single estuaries, and species which show a different variety on each estuary or group of neighboring estuaries." This may be another indication that the bay inside Nova Scotia and Georges Bank possibly had low salinity during part of the late-Quaternary epoch.

The northward inclination of the land during glaciation in southern New England at any given moment is, of course, unknown on account of the gradual uncovering from the ice and the unequal rising of the land at the different points. But the highest traces of the sea on the east coast and of the narrow lakes

in the Hudson and the Connecticut valleys are determined at a number of places (Woodworth, 1905; Emerson, 1898; Fairchild, 1914, 1919; Goldthwait, 1922, etc. Fairchild sometimes puts the levels higher than other geologists; moreover, he thinks that the waters in the valleys formed bays of the sea). These water levels in the lower parts of the valleys rise northward at the rate of several feet per mile. Because the land was rising while one place after another was uncovered from the ice and exposed to the attack of the sea, the upper marine limit that we see slants more gently than the actual tilt.

The chief later events relating to the level have been upwarping of land (strongest in the north), eventual sinking of part of the continental shelf, and rise of the sea level. These occurrences evidently were such as to lead to complicated results in the form of transgressions and regressions if the land and the sea rose at different rates at different times. Unequally rapid rise of sea level in respect to time is evident, as the rate of ice wastage varied considerably and the depletion occasionally was even less than the supply. The rise of the land perhaps took place spasmodically between long intervals of quietness, as seems to have been the case where the process is somewhat known. The actual amount of uplift at any point inside the isobase of zero equals the sum of the height of the marine limit and of the amount that the sea level stood lower than now when the limit was registered.

In the zone between Haverstraw and Albany, at any rate, the upheaval of land proceeded more rapidly than the rise of the sea level so that the land finally came to stand higher relative to the sea than at present. This is shown by the fact that the tributaries to the Hudson below Albany have much too deep and wide mouths in the unconsolidated glacial deposits (Woodworth, 1905, pp. 229–231). The Hockanum, a tributary of the Connecticut just south of Hartford, near its mouth flows in a too wide valley of erosion now in the process of growing together. The whole coastal region of New England has lain higher relative to the sea than at present, as shown by submerged stumps and peat bogs and drowned valleys. At Boston the shore line once

stood at least 20 feet (6 m.) lower than today (Johnson, 1925, p. 556). In Maine it may have lain much farther out than this. During the time that the shore lay outside the modern one, or at least during part of it, the land may have stood actually lower than today. Finally the land may or may not have been raised above the modern stand (Antevs, 1928a).

The shore line may have lain outside its present position for about one-half of the late-glacial and postglacial time, or until fairly recently. This is suggested by the fact, among other things, that the river mouths have not yet become filled out and by Shimer's (1918, p. 462) find at Boston of subfossil marine fauna which resembled that now off the Virginia coast and which occurred under conditions which show that the land then stood 16 to 18 feet (4.8 to 5.5 m.) higher above sea level than today and that it was in process of sinking. The deposit evidently dates from the postglacial temperature maximum, which may have prevailed from 7000 to 3500 or 2500 years ago (Sandegren, 1924, pp. 43, 52). Considering these conditions, fluctuations of sea level, and eventual superelevation of land, it appears probable that the existing relation of land and sea was at least nearly attained 6000 to 3000 years ago (Antevs, 1928a). Whether the shore line now is stable or whether transgression due to land sinking is still going on cannot be determined from conditions so far known. Stability is considered likely by Johnson (1925, p. 585). Land sinking of about one foot in a hundred years is assumed by various geologists and engineers.

In the region of Peekskill on the Hudson the transgression by the sea may have been followed by slight emergence, for at localities 110 and 111 a postglacial clay bed marking transgression is overlain by modern peat and sand (see p. 180). If this transgression is local and stability of land and sea is normal, the regression here is due to upheaval of land.

### Northern New England and Eastern Canada

During the uncovering from the ice of northern Vermont, lakes were ponded in valleys between the ice border in the north and

land in the south. In valleys which like the Connecticut Valley now drain southward the damming in the south was caused by northward warp of the earth's surface. In the Passumpsic Valley, the continuation of the Connecticut Valley straight northward, a lake extended to the vicinity of West Burke, 16 miles north of St. Johnsbury, where there are large quantities of gravel and sand. When the ice front had receded beyond the divide, other water bodies, given the general name of Glacial Lake Memphremagog, were dammed in depressions and valleys. Glacial Lake Memphremagog seems to have discharged first at Willoughby Lake and at Glover southward to the Passumpsic, later at Island Pond eastward to the upper Connecticut River, and still later at Elligo Pond southwestward to the Lamoille River (Hitchcock, 1910, p. 202). Subsequently, lower and lower outlets may have been opened on the Canadian side.

In the Champlain Basin there was a lake much wider than the modern Lake Champlain, which discharged southward through the Hudson Valley. It is called Glacial Lake Champlain or Lake Vermont. In the Winooski, the Lamoille, and other valleys also lakes were dammed, which finally were drained to and became parts of Lake Vermont (Hitchcock, 1906a; Merwin, 1908; Gordon, 1926, p. 290). In the year 7294, as the pass at 740 feet altitude 3 miles northeast of Stowe became ice free, the lake at Hardwick-Morrisville in the Lamoille Valley coalesced with that in the Winooski Valley, as shown by the sediments at localities 150 and 151 (cf. p. 192). The drainage had no effect upon the sedimentation in the Connecticut Valley (Antevs, 1922, Pl. 5). After the year 7330 Lake Winooski was lowered somewhat through the valley of Huntington River and Hollow Brook (cf. p. 122). In the year Essex 181 a very important tributary to the Lamoille Bay at Cambridge was discontinued, as shown by the abrupt change at localities 166 and 168 from coarse silt to stiff clay. In the year Essex 303 a lowering of the level of Lake Vermont seems to have taken place (cf. p. 235). In the year Fairfax 139 either a new tributary began to discharge into the Lamoille Bay or Lake Vermont underwent a lowering (cf. p. 236).

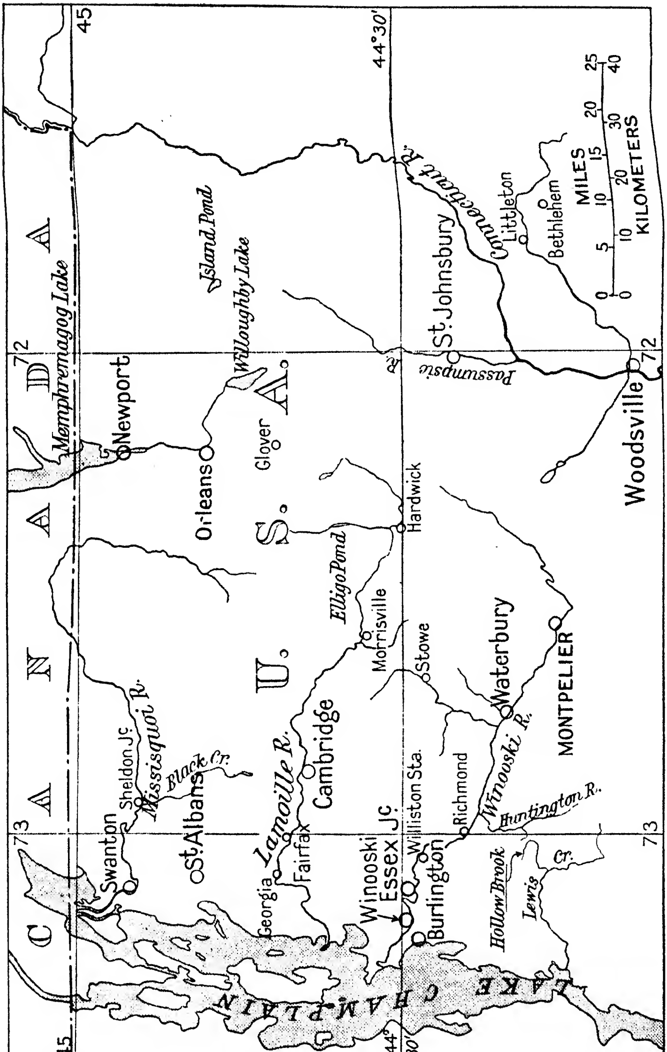



Fig. 2—Northern Vermont.

The supposed uppermost levels of Lakes Vermont, Winooski, and Lamoille have been determined at a number of places, but the figures do not wholly agree (Woodworth, 1905, Pl. 28; Hitchcock, 1906a, 1910; Merwin, 1908; Fairchild, 1916, p. 6; 1919, Pl. 5). Lake Vermont reached altitudes of about 675 feet at Burlington, 775 feet at the International Boundary on the east side of Lake Champlain, and 740 feet at the frontier on the west side.

When the ice border reached the northern side of Covey Hill just north of the International Boundary and west of Lake Champlain, Lake Iroquois, which formed the first important stage in the Ontario Basin after its water body had become separated from that of the Erie Basin, was drained into Lake Vermont. The lowering was from about 1030 feet to 740 feet, and thus amounted to about 290 feet (Fairchild, 1919, Pl. 5). This event may have had little effect on Lake Vermont except possibly to cause slight rise of the level. The resulting lake west of Covey Hill is called Lake Frontenac. Lake Frontenac-Vermont discharged southward by way of the Hudson. It came to an end by drainage to the sea through the St. Lawrence Valley—drainage which probably took place across the thinned ice long before the ice border had retired to the vicinity of the city of Quebec, when it stood between the frontier and Montreal (Antevs, 1925b, p. 64; Fig. 27, p. 74). The drop in level of the water may have amounted to more than 217 feet, the sea then probably standing somewhat below the marine limit at Covey Hill, which, according to Goldthwait (1913b, p. 125) lies at 523 feet. As the land stood several hundred feet lower in relation to sea level than at present, a bay of the sea then overspread a large part of the Champlain-St. Lawrence-Ottawa lowland, which is called the Champlain Sea. Marine molluscs migrated southward in the Champlain Sea to Crown Point station, somewhat south of Lake Champlain, and toward Lake Ontario as far as Brockville, and westward 60 miles beyond Ottawa to Fort Coulonge (Goldring, 1921, pp. 165, 185, map 1), and northward to the vicinity of Venosta, 34 miles north of Ottawa. All these extreme limits may have been reached later, that is, during the second deep-water stage, as only *Port-*

*landia arctica* is definitely known to have inhabited the Champlain Sea during its first stage. Marine organisms have been found in Vermont up to nearly 300 feet of altitude at Burlington and almost 400 feet at St. Albans (Hitchcock, 1910, p. 207), that is to say practically up to the levels regarded by Merwin (1908, p. 125) as the uppermost traces of the Champlain Sea.

After the fall of Lake Frontenac-Vermont to sea level the Winooski and the Lamoille Rivers began in earnest to transport into the sea the thick and extensive glacifluvial deposits that had been laid down in their valleys, thus building up the enormous deltas at Essex Junction-Winooski and at Georgia (Hitchcock, 1906, 1906a, 1910). As neither deposits nor shore features in Vermont have been given careful study, the history of the Champlain stage here is little known. It seems highly probable, however, that it was fundamentally the same as in the Montreal-Ottawa region, that it was divided by crustal oscillations into two deep-water stages and one intervening shallow-water stage (Antevs, 1925b, p. 73).

Conditions in southeastern Canada during the waning of the last ice sheet, as worked out both from old and from new studies, have been recently treated (Antevs, 1925b, pp. 59–79), so that here only the chief events will be summarized and the treatment brought up to date. To the published evidence of three marine late-glacial stages much new evidence was added by Professor J. W. Goldthwait and the writer in 1925 (Goldthwait, 1926).

Two beds of marine clay separated by a break in deposition, erosion, and usually by a bed of gravel or sand were observed at a number of places.

The most important places are as follows:

*A.*—Eight miles northeast of Ottawa on the Blanche River, a mile northwest of its junction with the Ottawa River, two beds of marine clay separated by sand were exposed.

*B.*—East of Kirks Ferry, on the Blackburn Brook, 300 yards from its mouth on the Gatineau and ten miles north-northwest of Ottawa, the following section was exposed by a slide:

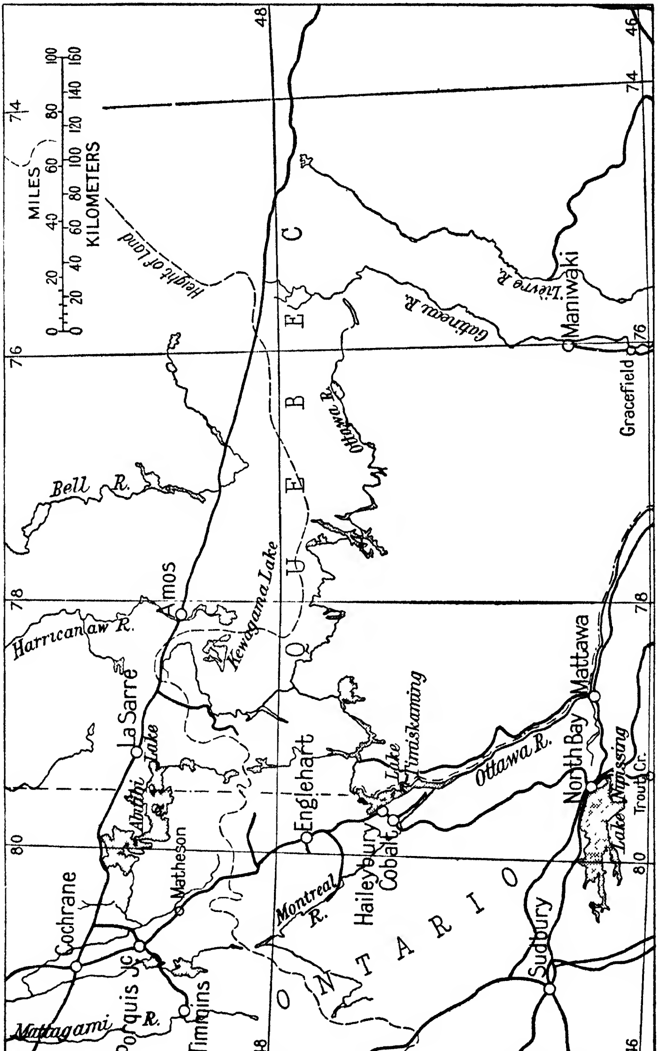



FIG. 3—Parts of northern Ontario and Quebec.

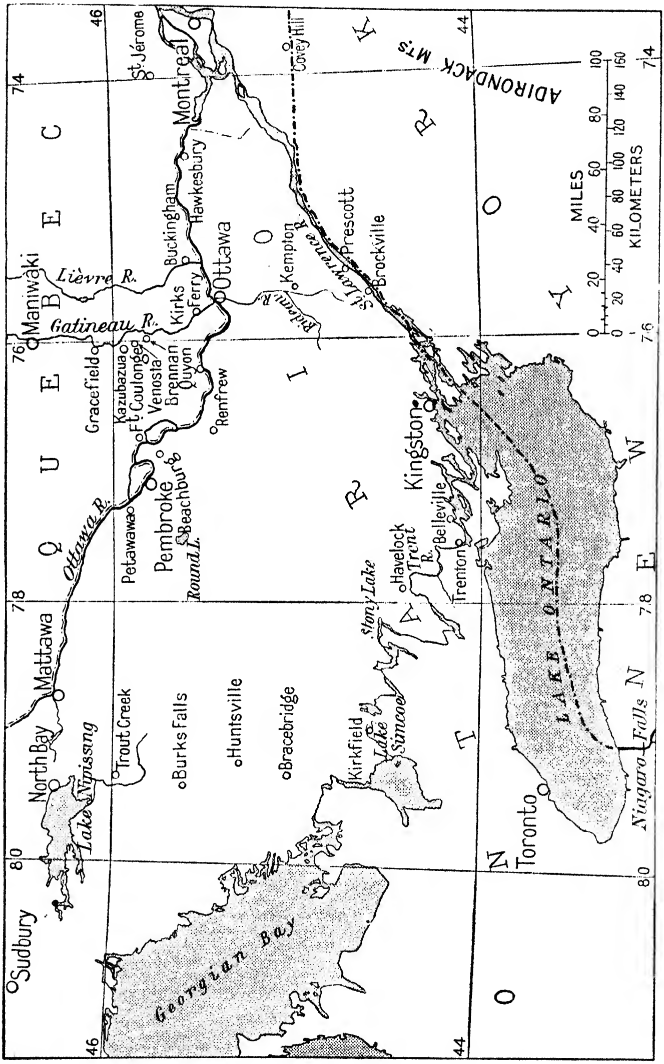



FIG. 4—Parts of southern Ontario and Quebec.

50 feet marine clay without shells. The 20 topmost feet consist of alternating layers of sand and clay. Varvity indistinct. The 30 lower feet consist of stiff, almost massive, clay.
Sharp limit.
10 feet clay and sand. Shells of *Saxicava* abundant, especially at top. *Macoma* sp. rare.
10 feet covered.
3 feet stiff marine clay with *Portlandia arctica.*
Brook level.

*C.*—Twenty-eight miles north-northwest of Ottawa, and 400 yards north of Brennan station, a road cut showed:

Several feet of clay, massive or with faint varves ½ to 1 inch thick. Near base several shells of *Portlandia arctica* and one shell of *Macoma* sp.
Sharp limit.
0–3 feet gravel with cobbles.
Erosion surface.
Several feet of sand and clay in alternating layers but not in varves No shells observed.

The altitude of the disconformity between the lower and the upper clay is 370 feet.

*D.*—Twenty-seven miles west-northwest of Ottawa, at the flour mill in Quyon, the following section was observed in the river bank and in a sand pit 50 yards from the river:

0–4 feet stiff, dark-brown, marine clay. No shells observed.
Erosion surface.
5 feet current bedded sand.
Erosion level.
0–8 inches gray, clayey silt with *Saxicava arctica* and *Macoma* sp.
12 feet yellow sand.
Contact not well exposed, but coarse sand just above and stiff clay just beneath.
20 feet marine clay, stiff, dark-gray, with very indistinct varvity. About 4 feet above the base numerous large shells of *Portlandia arctica.* The upper 10 feet of the clay are covered and so are a few feet at the bottom; but there cannot be any fresh-water clay at the bottom, as flat bed rock is exposed in the river a few yards from the shelly clay.

*E.*—Twenty miles east-southeast of Pembroke, at Forester Falls, at a brickyard in the southern edge of the village this section was exposed:

3 feet clay with indistinct varvity. The lowermost varve traced for 8 feet.
4–8 inches silty sand.
Uneven erosion surface.
6 feet massive clay, apparently marine, though no shells were observed.

The erosion surface lies at an elevation of 430 feet.

*F.*—Locality 161, Pembroke, shows the three marine stages (see p. 220).

*G.*—The clay pit of the brickyard, on the river 2 miles south of Pembroke, presents the following sections:

A. Near road.
 5 feet light-brown, indistinctly varved clay.
 1 foot fine sand.
 1 foot sandy clay.
 3 feet very stiff, brown, nearly massive clay.
B. 25 yards west of A.
 5 feet stiff, indistinctly varved clay.
 Sharp limit.
 ½–2 feet fine gravel, that could be traced for 12 yards.
 Uneven erosion surface at 380 feet altitude.
 ½–1½ feet sandy clay.
 1 foot stiff, brown, nearly massive clay.

From these observations and those previously described (Antevs, 1925b) the changes of level in the Ottawa region appear to be as follows. During the first marine deep-water stage, following directly after the drainage of Lake Frontenac to sea level, the highest shore lines in the region north of the Ottawa River were registered, for instance the 690-foot (210 m.) limit at Kingsmere, 8 miles northwest of Ottawa. Subsequently land upheaval caused the sea to fall to the bottom of the Ottawa Valley, to the 250- or 200-foot level or possibly still lower. After a slight submergence and an insignificant uplift a considerable land sinking took place, culminating in the second marine deep-water stage. In the Ottawa region the shore line reached at least the 470-foot

level. Then came another uplift, during which the shore receded at least below the level of 270 feet. Subsequently slight sinking set in followed by uplift. The rise proceeded until the shore line reached the level of about 250 feet. It has been supposed that a body of fresh water called Lake Ottawa now became isolated in the Ottawa Valley west of Hawkesbury (W. A. Johnston, 1916a, p. 7; Antevs, 1925b, p. 73), but observations by Goldthwait and the writer make it seem probable that the water sheet was a marine estuary.

These strong oscillations of the shore line seem mainly to be due to vertical movements of the earth's crust, as only a slight rise of sea level can have taken place during this time of insignificant ice melting. The limit levels are yet little known, but other material bearing upon them will before long be presented by Goldthwait in publications of the Geological Survey of Canada.

The basins of Lakes Huron, Michigan, and Superior in late-glacial time were occupied by Lake Algonquin. During its early stages Lake Algonquin discharged at Port Huron and at Chicago. When the ice border had receded to Stony Lake, 88 miles northeast of Toronto, the Trent Valley offered a lower outlet to Lake Ontario. This outlet soon took the whole overflow, inaugurating the Kirkfield stage, which lasted for a long time. It came to an end as a result of the rise of the Trent Valley, whereupon the old outlets at Port Huron and Chicago, which had remained stationary, were again adopted. The Trenton region finally came to stand higher in relation to the water body in the Ontario basin than it does at present to Lake Ontario, for the base of the channel of the Algonquin River is submerged beneath the level of Lake Ontario (Coleman, 1922, pp. 31, 51). The uplift may have been the one that separated the early and the late parts of the Champlain stage. The oscillation may be recorded by a sand layer reported by Keele (1924, pp. 72, 73) in the clay at Belleville on Lake Ontario and at Foxboro, six miles north of Belleville. The Kirkfield stage may have corresponded to Lake Iroquois, Lake Frontenac, and the first marine deep-water stage in the Ontario basin and the St. Lawrence lowland.

The Port Huron channel was quickly cut down through clay, so that it took the whole discharge. The resulting stage lasted until the great upheaval of the northern part of the Great Lakes region in late Algonquin time had nearly been completed, and the thinned ice barrier in the Mattawa Valley gave way, letting the water level drop suddenly 25 or 75 feet. The uplift may have been that which made the Champlain Sea for the last time fall to the bottom of the Ottawa Valley. After the uplift had ended and the water level had adjusted itself to the new outlet past North Bay by way of the Mattawa and Ottawa Rivers, the stage called the Nipissing Great Lakes came into existence. It persisted until 3500 to 3000 years ago when uplift of the North Bay region for the last time shifted the outlet of the Upper Great Lakes to Port Huron.

Because of the northward inclination of the land due to greater suppression of the central parts of the glaciated area, an extensive water body overspread the Timiskaming lowland as it became ice-free. This is called Glacial Lake Timiskaming or Lake Barlow. It discharged by way of the Ottawa River and extended northward across the height of land in western Quebec.

North of the watershed, directly north of Lake Huron, another water body, Lake Ojibway, was formed. This discharged southward to the Nipissing Great Lakes. After the highest region south of Lake Abitibi had become uncovered, Lakes Barlow and Ojibway merged with each other. The resulting lake finally had an enormous extent. It reached from the southern end of Lake Timiskaming up to the Transcontinental Railway, and from Bell River (48° 20′ N., 77° 15′ W.) westward to the region north of Lake Nipigon (88° W.). It underwent many changes (see Antevs, 1925b, p. 75). The distribution of the varved clay, which will be presently treated, shows that Lake Barlow-Ojibway or a subsequent Lake Ojibway was drained suddenly. The drainage is probably recorded at localities 126 and 127 by the increase in thickness shown in the topmost varves, the increase then being due to the shallowing. The emptying may thus have taken place during the years 2022 (or 2015) to 2027, and the

yellow unvarved clay silt that directly overlies the varved clay may have been formed after this event.

The exact position of the ice front at the time of this drainage is not easily determinable. The ice border first receded an unknown distance north of Cochrane, for varved clay correlated with the Timiskaming varve series was measured at localities 136 and 137. Then it readvanced to 33 miles south of Cochrane, to locality 127. As the boulder clay laid down by the ice during this advance normally forms the uppermost deposit, where not covered by peat or other distinctly postglacial sediments, it is evident that Lake Barlow-Ojibway was completely drained before retreat was renewed. South of the invaded area there does not seem to be any varved clay corresponding to the halt suggested by the large outwash deposits, nor any long series of thick varves at localities 124 and 125 which are situated relatively close to the large wash deposits north of Porquis Junc. The fact that in the new gorge of the Frederick House River, where the measurements 128–135 were made, boulder clay directly overlies stiff, thinly varved clay perhaps means that the drainage preceded the readvance of the ice here. Indeed, the fact that the drainage must have taken place to James Bay and probably occurred by overflow over the thinned ice and the cutting of a trench, makes it probable that it occurred when the ice border stood far north of Cochrane, and before the readvance. The drainage was complete, so that there may not have remained many more or larger water bodies than exist today. Some of the small basins have become drained, and the varved glacial clays exposed, as the rivers have cut down through them in postglacial time; but, because of the early drainage of Lake Barlow-Ojibway, geochronological study of the later ice recession is perhaps made impossible.

# CHAPTER VII

# RATE OF RECESSION OF THE LAST ICE SHEET FROM LONG ISLAND TO NORTHERN ONTARIO

The field within which systematic search has been made for clay sections and other significant records of recession reaches from the terminal moraine on Long Island and New Jersey northward over New England and New York State and across Ontario to a point midway between Lake Timiskaming and Hudson Bay. In this chapter, wherever hitherto unpublished material appears, it is presented in some detail; but where measurements have already been described and discussed in earlier papers the treatment here is summary. It is intended thus to show wherein the attempt to construct a continuous chain of exact recession measurements has succeeded, and where, for this reason or that, it has failed. Where links in the chain are missing, conjectural estimates are substituted for more exact figures.

## The Terminal Moraines

Because of lack of clay or of exposures no varve measurements have been made outside, on, or just inside the terminal moraine, but the most peripheral clay sections studied are that at Glen Head on Long Island, locality 122, and those in New Jersey (see p. 173).[1] The length of time of the formation of the terminal moraines and the rate of the ice retreat between them, therefore, cannot be directly determined.

During its whole existence the ice sheet received nourishment and suffered wastage. When nourishment was dominant, the ice shield expanded; and when wastage was the greater, it shrank. The transition from expansion to retreat marks a time when nourishment and wastage counterbalanced each other and after

[1] Gerard De Geer's method of studying the ice retreat by means of the varved glacial clays is described in earlier papers by the present writer (1922, p. 3; 1925b, p. 9).

which depletion became dominant; it marks the most important climatic change in late-Quaternary time. If the amount of débris in the ice and the rate of forward motion of the ice were known, the length of time of formation of a certain amount of drift could be calculated. As it is, at best a rough estimate can be made. The most accurate way may be to compare the terminal moraines with other morainic lines whose time factor is known. The best norm is the Fenno-Scandian moraines and, above all, the most carefully studied parts of them, the Salpausselkäs in Finland.

The average width of the Salpausselkäs is about 1 ½ miles (2.5 km.) and the average height some 65 to 165 feet (20 to 50 m.) though the altitude locally is as much as 260 feet (80 m.). The material of the ridges consists of sand, gravel, cobbles, and rounded boulders (Leiviskä, 1920, p. 240). Till does not occur or, in any case, does not play much rôle. The outer ridge, or series of ridges, represents 225 years, the inner one 183 years, and the intervening recession 251 years, making in all 659 years (Sauramo, 1918, pp. 23, 35).[1] The halt of the ice edge at the outer morainic line may have been largely due to low summer temperature, for the clay varves then formed are very thin. However, the moraine consists of stratified drift indicating ice melting. The halt at the inner line, during the formation of which quite thick varves were deposited, probably was chiefly due to rich nourishment of the ice (Antevs, 1925b, p. 53). These halts occurred during a fairly late stage of the waning of the ice, when it must have become considerably thinned out at the center, and the rate of ice motion probably was small or similar to that of the North American ice sheet during its greatest extent. However, the débris load may have been essentially greater in Finland, because of concentration through long-continued surface ablation, so that the formation of equally large ridges of the same kind of drift may have taken a much longer time in Long Island than in Finland.

The morainic lines in Long Island are notable features of the

[1] Ramsay (1924, p. 28) holds that the conspicuous varve 0 indicates the year when the Baltic ice lake was drained, not the year when the ice border left the Inner Salpausselkä, and that consequently the time of the formation of the inner morainic ridge is not yet determined.

topography, rising in many places 100 to 150 feet (30 to 45 m.) above the surrounding level and having an average width of a mile or somewhat less (M. L. Fuller, 1914, p. 32). The outer moraine, the Ronkonkoma Moraine, consists prevailingly of gravel and sand, while the inner moraine, the Harbor Hill Moraine, carries much till, especially at the western end of the island, where it represents a marked advance of the ice edge. The Long Island moraines are thus both narrower and lower than the Salpausselkäs. On their distal sides, however, there are wide wash plains, so that the material in all may be larger than in the Salpausselkäs. The Ronkonkoma Moraine consequently may represent several hundred years.

After the upbuilding of the Ronkonkoma Moraine the ice border receded. When it had reached about the present northern shore of the island it halted and readvanced, forming the Harbor Hill Moraine. The distance between the two stationary positions ranges from zero in the western part of the island to some 12 miles (19 km.) in the eastern. The retreat may have taken several hundred years. Part of the material in the Harbor Hill Moraine, especially in the western part of the island, was pushed together by the ice as it advanced. The deposition of the stratified drift was fairly rapid, for the numerous kettles of the wash plains indicate quick burial of detached ice blocks (M. L. Fuller, 1914, p. 211). The oscillation of the ice border, even if it was repeated, need not have taken a long time. The moraine and associated outwash may represent a somewhat shorter time than the Ronkonkoma Moraine, and the whole morainic belt some 2000 years.

In New Jersey the terminal moraine is undivided. Its average width is about a mile, and its average thickness is probably somewhat less than a hundred feet (Salisbury, 1902, p. 93). In some parts till predominates, in other parts gravel (Salisbury, 1908, p. 15). Outside the moraine there is generally a belt of gravel and sand, but the amount of this outwash material is not large (Salisbury, 1902, Pl. 28). It is not known whether the moraine represents both morainic lines in Long Island or just the inner-

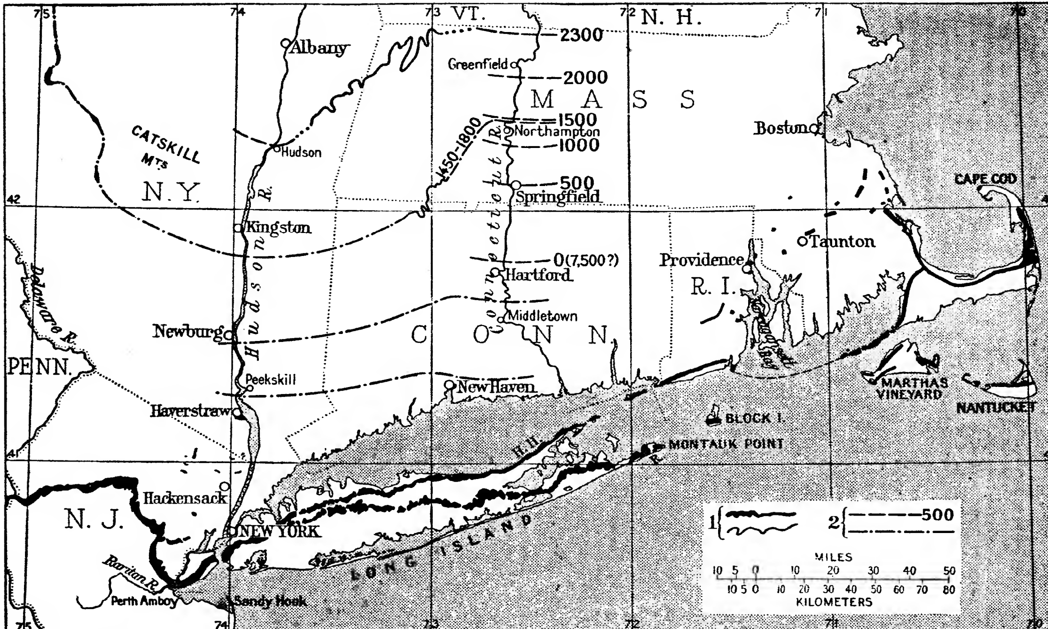



FIG. 5—Moraines (1) and a few positions of the last retreating ice border (2) in southern New England and adjacent regions. 7500? indicates the probable number of years taken by the halt at the outermost moraines and by recession from them. (Moraines from Salisbury, 1902, Pl. 28; Fuller, 1914; Woodworth, 1897; and F. B. Taylor, 1903. Ice borders chiefly from Antevs, 1922, Pl. 6.)

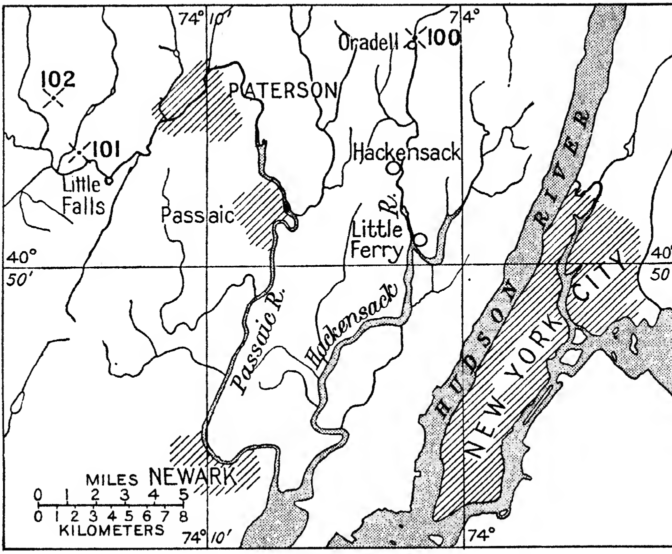
most one, with the correlative of the Ronkonkoma Moraine erased by advance of the ice front across it. At any rate the moraine may not represent a longer time than the whole morainic belt in Long Island.

### The Belt between the Terminal Moraine and the Line Newburg, N. Y.–Hartford, Conn.

The actual rate of the ice retreat in the peripheral zone of the glaciated area in North America and the time involved have not been determined on account of scarcity of clay and of exposures (see p. 169).

In the belt just inside the terminal moraine in New Jersey the recession of the ice border was probably fairly continuous, since stadial moraines are scarce. At localities 101 and 102, however, the ice edge oscillated, at 102 probably repeatedly. The time represented by these oscillations is surely more than 400 years. From our present knowledge it seems likely that the rate of retreat in the beginning was quite slow, slower between the terminal moraine and Hackensack than between that city and Haverstraw, N. Y. As the recession from Hackensack to Haverstraw may represent a minimum period of 1950 years (see below), that from the moraine to Hackensack, which is an equal distance, or 23 miles (37 km.), probably took some 2500 to 3000 years.

Although Lake Hackensack at a few places was connected with lakes in basins to the west (Salisbury, 1902, Pl. 28), no material worth mentioning appears to have been brought from them into the northern part of the Hackensack basin, where the varve measurements were made. The drainage area can be considered then to have comprised only the narrow tract between the Palisades on the Hudson and a line running about due north from Hackensack. As the level of the late-glacial lake over the Hudson at Haverstraw, N. Y., stood only about 100 feet above the present sea level (Woodworth, 1905, p. 100), while the lowest passes on the watershed west and southwest of Haverstraw now lie at altitudes of 480, 520, and 540 feet respectively, the northern limit of the Hackensack drainage area in late-glacial time lay at



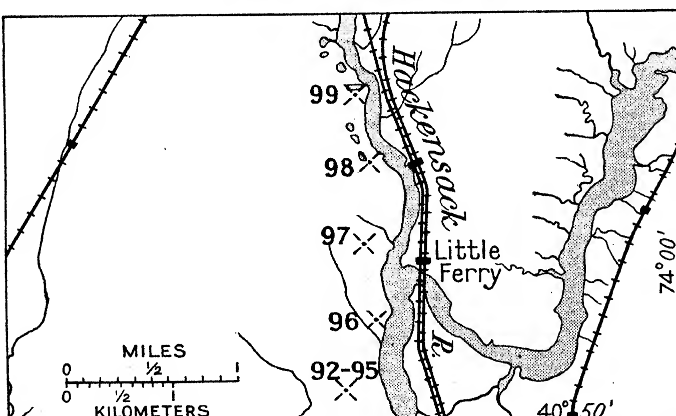




FIG. 6 (above)—Position of the localities examined in eastern New Jersey, west of New York City (localities 92–102).

FIG. 7 (below)—Detailed position of the localities at Little Ferry, N. J. (92–99).

its present position. Thus, when the ice border reached the north side of the high ridge at Haverstraw, the Hackensack area lost direct communication with the ice. This event apparently caused deposition of varved clay at Hackensack to cease, so that probably all beds except the uppermost nearly massive clay bed

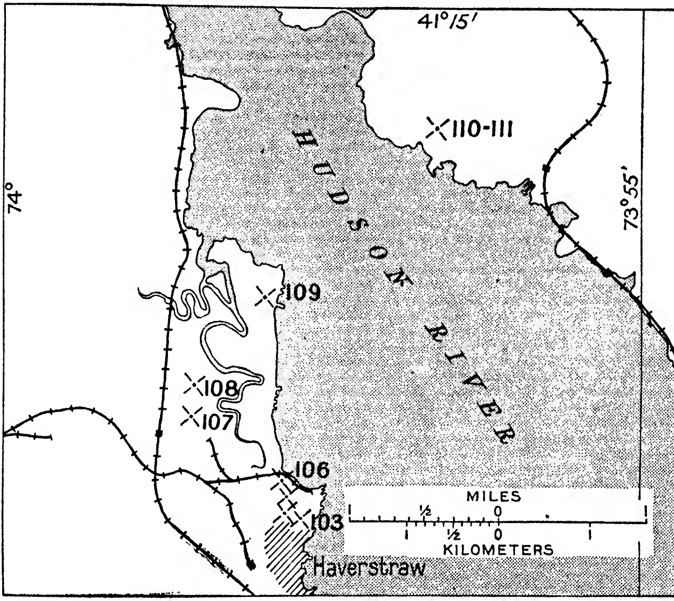



FIG. 8—Position of the localities examined in the region of Haverstraw, N. Y. (localities 103–111).

at localities 92, 95, and 96 and the top sand were formed previously. As the clay beds in the present exposures contain some 1950 varves (see locality 96, p. 174; cf. Reeds, 1924, 1925, 1926), the ice retreat from locality 99, where the substratum was reached, to Haverstraw may have taken at least as many years. It is possible that deposition of varved clay at locality 96, because of insufficient depth of water, was interrupted before the ice front reached Haverstraw, or that deposited varved clay was later

eroded away. As the distance from locality 99 to Haverstraw is 23 miles (37 km.) the rate of ice recession averaged at most 62 feet (19 m.) annually. The insignificant thickness of the varves even near the substratum also indicates small annual melting and retreat.

In the clays at Hackensack no traces of readvance and overriding by ice have been found. For the first 946 years after the uncovering of locality 99 the retreat may have been fairly uniform, judging from the even thickness of the varves. In year 946 the ice edge had probably reached about the 41st parallel. The changes in amount of material at varves 947 and 1101 are too sudden and too great to be caused by changes in the amount of the summer ice melting. If the thin layers in the zone beginning with varve 953 be varves, the changes could be readily explained by a sudden increase of the drainage-area year 947 and diversion of the mud-carrying current into another basin during the years 953 to 1100; but it is not evident from the physiography of the region how these latter events could take place. Perhaps drift and buried ice blocks formed temporary barriers inside of which part or most of the glacier mud was trapped, and the varves 947 and 1101 mark sudden cuttings through of the barriers. There are near the state boundary particularly two belts of recessional moraines (Salisbury, 1902, p. 517, Pl. 28; Woodworth, 1905, p. 94).

The gradual decrease and the relative evenness of the thickness of the varves from number 1101 and upwards suggest a fairly uniform ice retreat. The little moraine just south of Haverstraw (Woodworth, 1905, p. 98) probably does not represent a long time.

The following profiles were shown to the writer in 1924 by the late Dr. H. H. Robinson of Yale University in the shore bluff (exactly at 73rd meridian) at Woodmont, seven miles southwest of New Haven, Conn.

A.

Topmost, 2–3 feet yellow-brown clayey till.

5–7 feet yellow-brown silty sand with scattered boulders. Probably deposited below the ice, though not typical till.

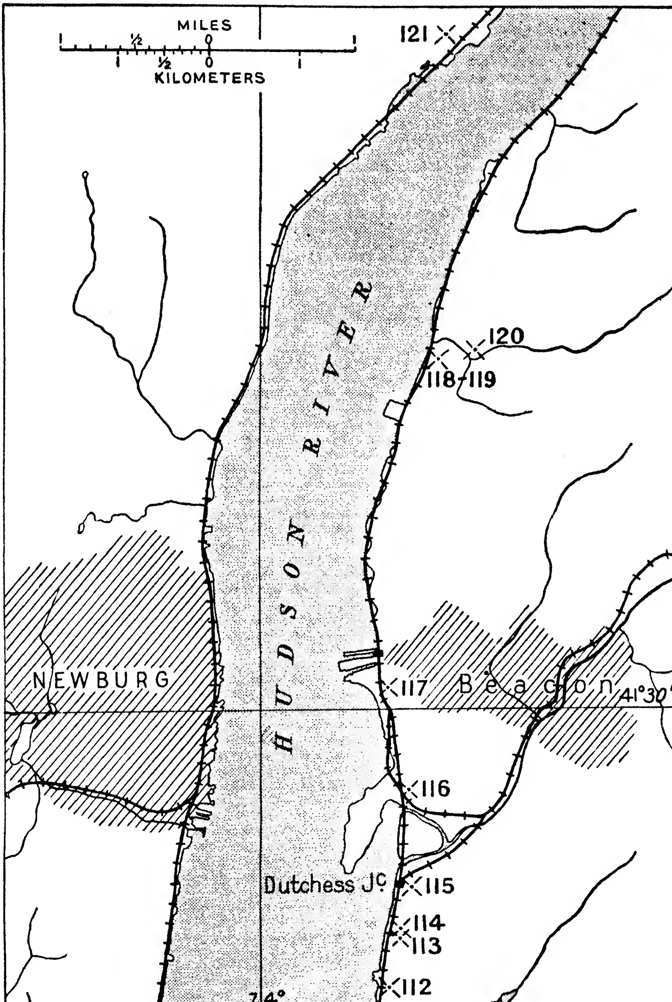



FIG. 9—Position of the localities examined in the region of Newburg N. Y. (localities 112–121).

(These two top beds locally represented by a single bed of silty and clayey till.)
8 inches red, clayey sand.
2 feet gray shingle—probably a beach.
2 feet gray till with concentrated large boulders.
Erosion level.
4–7 feet gray-green, very stony boulder clay or clayey till.
Bed rock at sea level.

B. 50 yards from section A.
Topmost, 12 feet yellow-brown clayey till.
Disconformity.
7 feet gray-green clayey till.
Sea level. Bed rock close below.

Profile A records ice over the place at three different times. Before the third advance the exposed part of the then ground surface lay in the shore line, as is indicated by the probable beach gravel and the concentrated boulders in the remains of the second till bed. The material may be quite local. A greenish schist occurs in the immediate vicinity, light gray gneiss some ten miles distant to the north, and red colored rocks a few miles to the northeast.

There was no way of dating the beds; and it is uncertain whether all of them originated during the last glaciation. While, therefore, no conclusions regarding the late-glacial events can as yet be drawn from the profiles, they well deserve description.

The varve series measured at Haverstraw is connected with that obtained near New Haven, Conn., which begins with varve number 180 at Haverstraw. Similarly the varve series from the region of Newburg, N. Y., is correlated with the previously published series from Hartford, Conn. (Antevs, 1922, Pl. 1), and extends this one back from varve 3000 to 2700. The series of 730 varves from Haverstraw cannot be connected with the measurements at Newburg and thus may not record ice retreat as far northward as that city. At neither place was bottom reached. The distance between localities 108 and 115, where the measurements extend nearest to the bottom, is 18½ miles (29.8 km.). The average rate of the ice retreat between the two points may have been less than 134 feet (41 m.) a year.

In the contemporaneous clays at Haverstraw and New Haven no indications of readvances of the ice edge have been observed. The great thickness of the varves at Haverstraw cannot be taken as absolute evidence of great summer melting and rapid recession, for the area drained to the Hudson at and above Haverstraw is

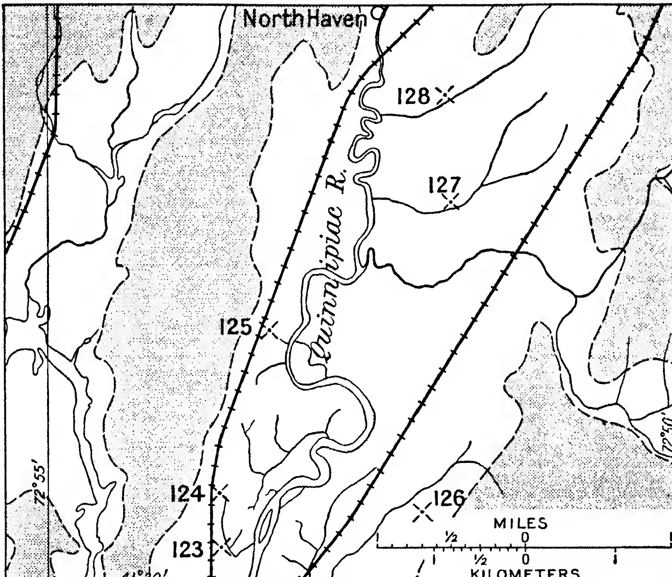



FIG. 10—Position of the localities examined on the Quinnipiac, northeast of New Haven, Conn. (localities 123–128). Locality 123 lies three miles northeast of Union station in New Haven. (Glacial lakes after F. Ward, 1920, Pl. 1.)

In this figure and in Fig. 11 the areas covered by late glacial lakes have been left white in accordance with Figs. 5 to 13 in Antevs, 1922, pp. 37–45, but in distinction from other figures in the present book.

wide, whereas the late-glacial basin of sedimentation was not much wider than the present Hudson. However, the varves at New Haven are also fairly thick, although the drainage area here was not broad in relation to the basin of deposition, so that it seems certain that the ice retreat in the belt north of Haverstraw and New Haven was much more rapid than in New Jersey.

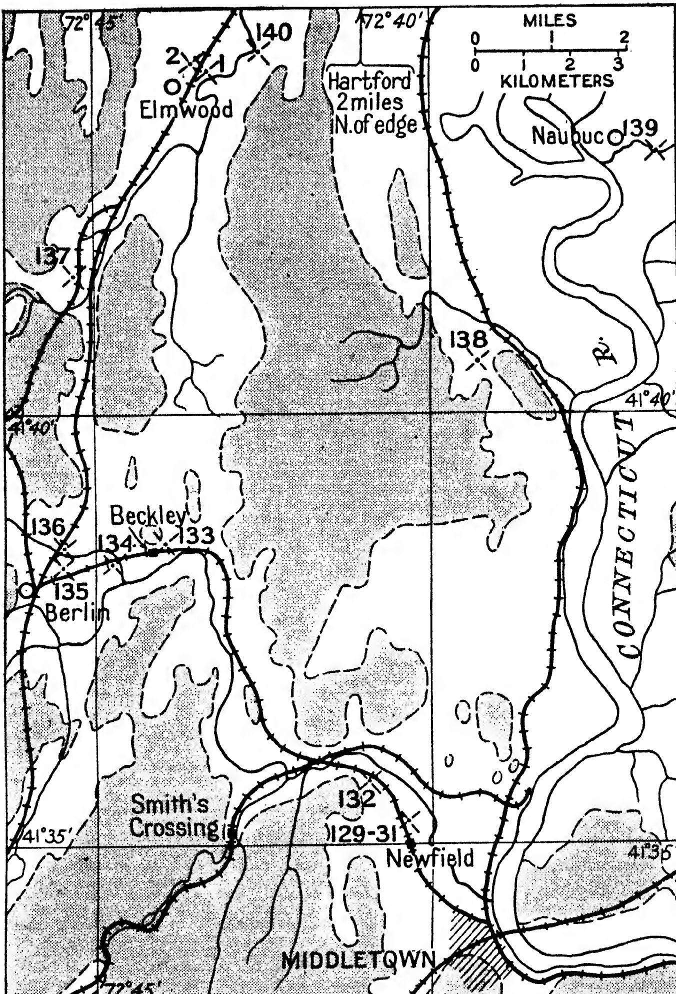



FIG. 11—Position of the localities examined in the region south of Hartford, Conn. (localities 1 and 2, Antevs, 1922, and 129–140).
Glacial lake sediments left white as in Fig. 10; after Loughlin, 1905, Pl. 1.

Between Haverstraw and Newburg no stadial moraines seem to have been found (Woodworth, 1905); but halts and readvances are recorded in the northern part of the corresponding zone in Connecticut, viz. between Middletown and Hartford. At all the studied sections between Middletown and Berlin and at locality 138 halfway between Middletown and Hartford crumpling and till kneaded into the clay show advances of the ice edge. Also at the brickyard on the railway, 1⅓ miles northwest of Middletown, and at Smith's Crossing, 3½ miles west-northwest of Middletown, the clay is disturbed and covered by till. The clay deposits in this belt contain only small numbers of quite thick varves, so that individual oscillations were probably of brief duration. Exceptionally slow recession is not suggested by any known conditions and is not probable, since the average rate between Hartford and Springfield amounted to 243 feet (74 m.) annually (Antevs, 1922, p. 76). Therefore, while the uncovering of the belt from Haverstraw to Newburg seems to have taken more than 730 years it may have taken somewhat longer but probably less than 1000 years.

In the Narragansett Bay region a number of halts in the retreat of the ice edge on lines about a mile apart are indicated by wash plains with steep proximal slope, showing that the material was piled up against a stationary ice wall (Woodworth, 1896). The large quantity of material in the plains shows that ice melting nevertheless was considerable. The halts may therefore have been largely due to excessive ice supply. Between the wash plains there is little sand and gravel owing perhaps to relatively rapid ice retreat or to the opening of fissures inside the ice margin in which the outwash was deposited. The regular recurrence of halts suggests that the former alternative most likely was the case and that the conditions were caused by climatic periodicity then chiefly fluctuations in nourishment of the ice. The length of this supposed period, so far as can be judged, is probably over 100 years. The Queens River Moraine to the west of Narragansett Bay and two morainic lines with wash plains in southeastern Massachusetts mark local or general readvances of the ice front

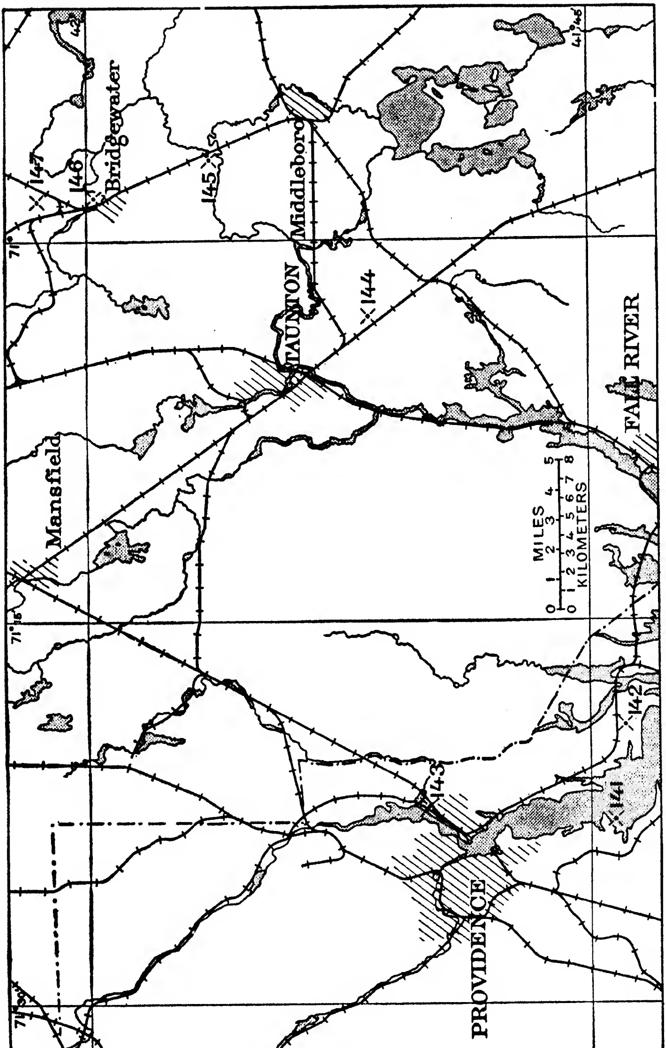



FIG. 12—Position of the localities examined in Rhode Island and Massachusetts (localities 141–147).

(Woodworth and Marbut, 1896a; Woodworth, 1897). The northern and most distinct of these lines, that running from Providence through Taunton and Bridgewater, may correspond to the ice advance at Middletown, Conn. Disturbances in the upper part of the clay and occurrence of boulders on its surface at Barrington, our locality 142, as observed by Sayles (1919, p. 29), indicate either advance of the ice and overriding of the clay or grounding of icebergs. Advance would probably not have taken place until at least a few hundred years had elapsed after the first uncovering. The surface beds at localities 145, 146, and 147 in southeastern Massachusetts are difficult to classify, but they are probably till deposited during advances of the ice border. Oscillations are also suggested by more or less thoroughgoing disturbances of the clay. The return to localities 145 and 146 then took place more than 90 and more than 200 years respectively after the original uncovering.

The terminal moraines, therefore, are thought to represent some 2000 years, and the ice retreat from them to the line Newburg-Hartford about 5500 years.

### Northern New England and Adjacent Quebec

In the region of Woodsville, N. H., the retreat of the ice border reached the highest rate observed in North America, amounting to 1100 feet (335 m.) a year (Antevs, 1922, p. 83). Between Woodsville and the junction of the Connecticut and Passumpsic Rivers the rate fell to 830 feet (253 m.) annually. South of St. Johnsbury the recession was still rapid, for locality 91 (Passumpsic) was uncovered not later than year 7060 and probably about 7050, or some 25 years later than Inwood, three miles to the south. The actual rate of retreat in the vicinity of St. Johnsbury is not known, for the clay is disturbed and the substratum has not been observed anywhere with certainty; but locality 172, one mile west of St. Johnsbury, may have been uncovered before year 7250. (Fig. 2, p. 95).

The formation of the alternating clay and gravel beds at locality 171 in the southern edge of St. Johnsbury (see p. 199), seems to prove that the ice edge stood in the vicinity for at least 200 years.

Morainal deposits at St. Johnsbury also indicate halt and readvance. Contemporaneously the ice front may have advanced in the Connecticut Valley nearly to the junction with the Passumpsic, for varve 7306 is bottom varve at locality 85 and the layers are absolutely undisturbed. The ice then uncovered locality 85 for

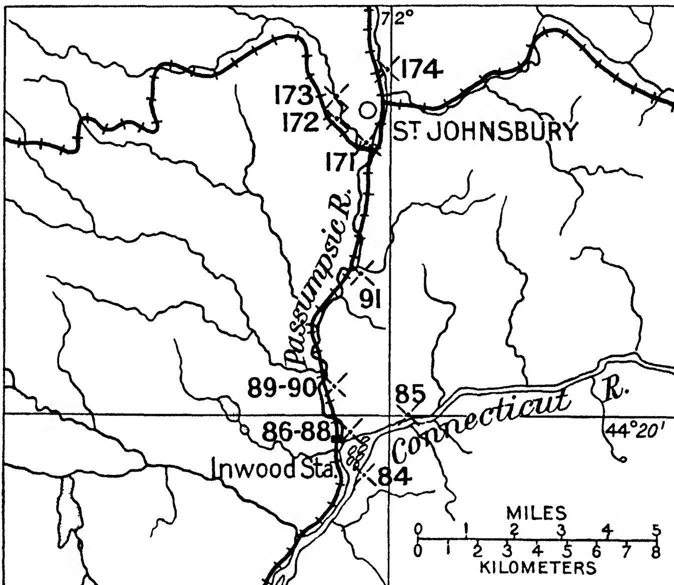



FIG. 13—Position of the localities examined in the region of St. Johnsbury, Vt. (localities 84–91, Antevs 1922, and 171–174).

the second time, some 280 years after it first uncovered it. The moraines at Bethlehem and Littleton, N. H. (Goldthwait, 1916), perhaps record the same oscillation.

With the exception of profile 171 the clay deposits at St. Johnsbury give no clue as to the rate of the ice recession (see p. 200). They only show, as does the clay at Inwood, that for some 800 years, or at least till year 7800, glacifluvial material was discharged into the lake of the Passumpsic Valley. The increase

in the thickness of the varves beginning year 7401 was probably to some extent due to increased melting (cf. p. 231).

In eastern Vermont the ice border may have retreated with a practically east-westerly front, judging from the straight north-southward trend of the striae (see Goldthwait, 1922). In the western part of the state the ice edge, because of rapid flow in the Champlain Valley, trended somewhat north of east to northeast.

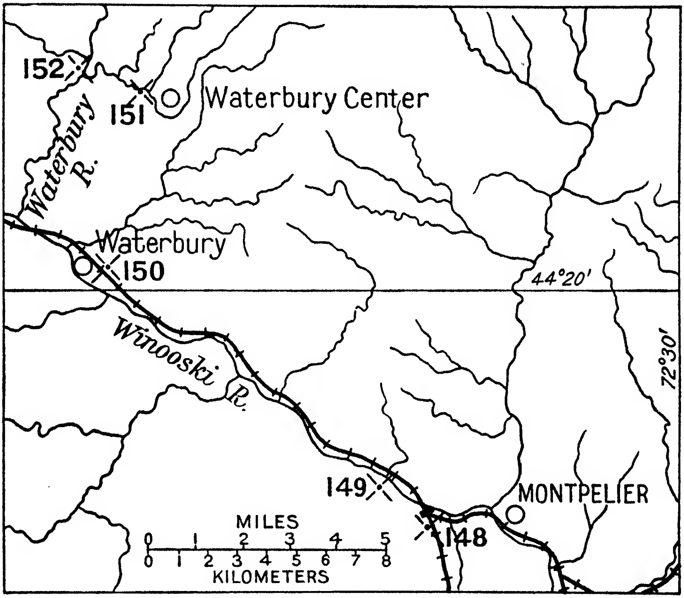



FIG. 14—Position of the localities examined in the Montpelier-Waterbury region, Vt. (localities 148–152).

The actual rate of the recession contemporaneous to the uncovering of the Inwood-St. Johnsbury region has not been determined elsewhere, but it evidently was rapid also in the Montpelier-Waterbury region, for the clay series at localities 148 and 151, though their lower parts are disturbed, begin with about the same varve.

But soon after the uncovering of locality 151, Waterbury Center, which took place before year 7066, the rate became slower. Not before the year 7294 did the ice edge reach the divide between the Winooski and the Lamoille River systems three miles northeast of Stowe, the opening of the passage being recorded by exceptionally thick varves at Waterbury (see pp. 94, 192). The watershed at 675 feet between Huntington River and Hollow Brook, 16 miles southeast of Burlington and 14 miles due west of Waterbury, may not have been uncovered until long after year 7330, for the stiff top clay and probably also the covering sand and gravel at localities 150 and 151, reaching heights of at least 650 feet, must have been deposited previous to the opening of this passage. The gorge-like valley is certainly narrow, being only some 300 yards wide at the 700-foot curve and 500 yards wide between the 800-foot curves, but the amount of water passing through it may not have been great enough to dam up the lake level to 750 to 800 feet at Waterbury, the probable stand during the deposition of the stiff top clay there. The view here held is supported by the fact that farther northwest in the Winooski Valley, in the region south of Richmond, sediments have been found only up to an altitude of 650 feet (Merwin, 1908, p. 131), while between Hollow Brook and Williston village, four miles southeast of Essex Junction, there is no passage that could have served as an outlet channel for the lake dammed in the Winooski Valley. Thus, in western Vermont, as well as in eastern, the retreat was slow and perhaps even interrupted by readvance about years 7200 and 7300. The ice edge perhaps reached Essex about year 7500. The ice front still had a southwest-northeast trend shown by the direction of the striae and the varve connections.

The bottom series obtained at locality 167, Cambridge, Vt., cannot be surely correlated with the series at Essex but the locality may have been uncovered about 40 years after Essex Center where the bottom of the clay was reached. Localities 164 to 170 in the Cambridge region may all have become ice-free at nearly the same time. The relatively great thickness of the

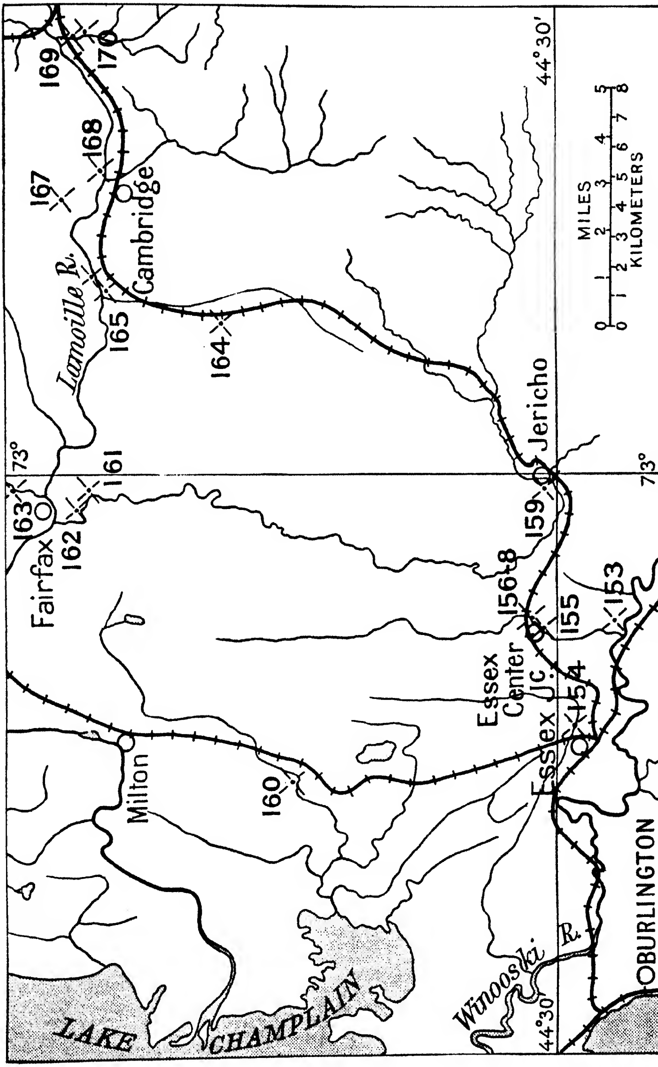



FIG. 15—Position of the localities examined in northwestern Vermont (localities 153–170)

varves may be no indication of considerable melting and rapid retreat, since the measurements were made in valleys or at the mouths of long valleys through which large areas were drained. The facts that the clay measurements at Cambridge and at Fairfax, only five miles apart, cannot be connected with each other, that locality 160 cannot be connected with either, and that the clay deposits may at least largely have been formed at different times, suggests very slow retreat or halt and readvance perhaps repeated several times in the belt northwest of Essex-Cambridge. Fairfax seems to have become ice-free at least 300 years later than Cambridge. Readvance may be recorded at locality 154, Essex Junction, by the thorough contortion and the lenses of till kneaded into the clay. The ice front did not advance to Essex Center nor to locality 153. Undisturbed parts of clay correlated with the normal curve of the region show that the oscillation took place more than 140 years after the first uncovering of Essex Center. Very little melting and perhaps halt and readvance is also indicated by the thin varves 49–79 and 101–138 at Fairfax. The insignificant thickness of the varves in the upper part of section 160 does not necessarily mean little melting but may be due, among other things, to great distance from the mouth of a glacial river.

In northwestern Vermont, north of the Lamoille River, glacifluvial deposits are scarce. In the Black Creek Valley there is between Fletcher and Fairfield no clay worth mentioning and north of Fairfield only little clay in the lowermost depressions, so that no sections were observed. On the Missisquoi River, east of Sheldon Junction, clay and sand are lacking, the river bed being cut down in till. Farther west, in the region of Swanton, the exposed sediments, which are more than 30 feet thick, are marine and shell-bearing. No exposure down to substratum was observed, so that the deposits may be underlain by fresh-water sediments. If so these cannot be thick. West of St. Albans near and on Lake Champlain thinly varved clays are to be found, but the deposits are small and at most ten feet thick. Clay was observed only at and slightly above the lake level. The scanti-

ness of glacial fresh-water deposits may indicate that the region was covered by a lobe of the ice until or till shortly before the beginning of the marine stage, which, as has been stated, came into existence some time after the ice edge had left Covey Hill. (Fig. 2, p. 95).

The time occupied by the retreat of the ice front from St. Johnsbury to the northern side of the divide cannot yet be determined, as the melt water even after this event was drained southward, so that little or no change was effected on the glaciaqueous deposition in the St. Johnsbury region; the deposition of clay at Inwood ceased because the basin began to fill with sediments. Morainic dams at both ends of Willoughby Lake (Jacobs, 1920, p. 283) indicate halts in the recession. Clay measurements make it evident that the retreat north of Willoughby Lake was slow. The series of 100 varves at locality 175 fails to conform with that at locality 178, five miles (8 km.) farther north, showing that the rate was less than 262 feet (80 m.) a year. The recession from locality 178 to locality 179, a distance of two miles (3.2 km.), took more than 25 years. Subsequent to the deposition of at least 31 varves at locality 179 the ice overrode the place after year Coventry 56, depositing a till bed on top of the clay. A little farther north the ice edge halted again, building up the moraine across the Memphremagog Basin on which the village of Newport is partly built. Localities 180 and 191 on opposite sides of Lake Memphremagog were uncovered at about the same time. The series of 140 varves obtained at these localities shows no agreement whatever with that at locality 144, Canada, 4 miles to the northeast. This means probably that the series actually do not overlap and that the ice edge halted in the intervening zone. However, the lack of correspondence may possibly be due to the rather different conditions of deposition, the Canadian locality occurring at a level 200 feet higher. Similarly the lack of agreement between the series 144 and 145 may not necessarily mean that the former was wholly deposited before the latter, as this is rather short and as they were obtained in valleys forming right angles to each other. Be this as it may, both series are older than that at

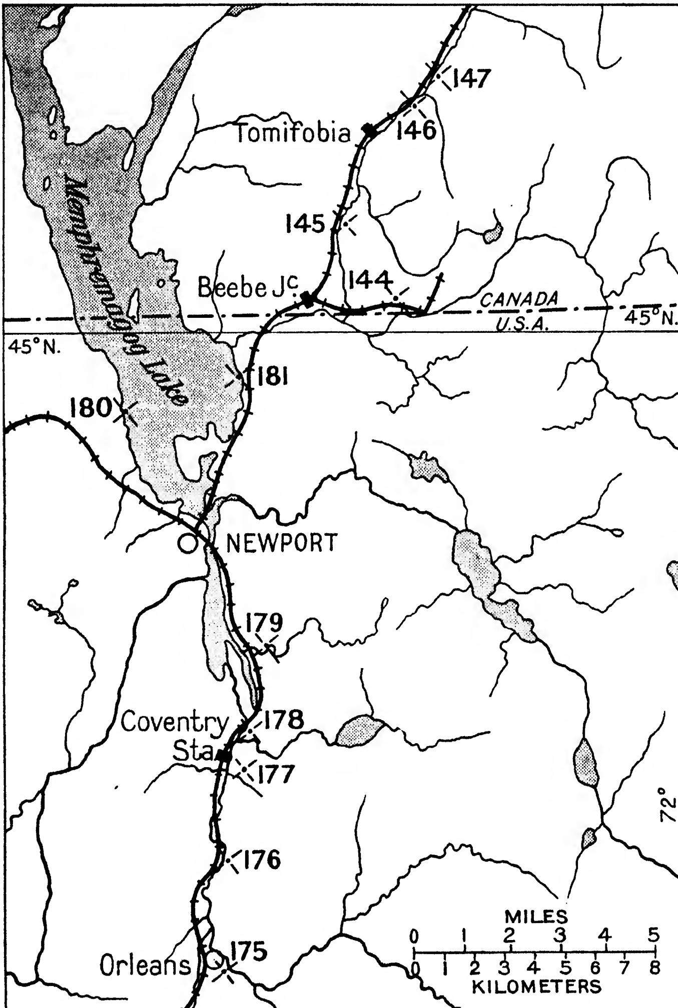



FIG. 16—Position of the localities examined in the region of Newport, Vt., and adjacent Quebec (localities U. S. A. 175–181 and Canada 144–147).

localities 146 and 147, where there is a sudden and remarkable increase in the deposition. As the varves below the increase consist of greasy clay and may occur high above the bottom, it does not follow that the ice retreat from point 144, a distance of four and a half miles, took more than 250 years.

Between locality 147 and the region of Sherbrooke no conditions giving any idea as to the rate of the ice retreat are known. The clay at the old brickyard two miles southeast of Lennoxville is rather thoroughly contorted, but no conclusive evidence of overriding by the ice was observed. The top bed at locality 148, Ascot, is perhaps till, marking a possible readvance some 25 years after the first uncovering. On the Clifton River, 16 miles east-southeast of Sherbrooke, and on the railway near Angus, 15 miles northeast of Sherbrooke, varved clay is overlain by till (Chalmers, 1898, p. 44; Keele, 1915, p. 93). At locality 149, Richmond, the clay is much disturbed and is covered by a heterogeneous mixture of silt and stiff clay which perhaps indicates overriding by the ice. Near the mouth of Rivière du Loup, 58 miles south-southeast of Quebec, two beds of clay and sand are separated by till, and the deposit is capped by boulder clay (Chalmers, 1898, p. 43; Keele, 1915, pp. 44, 93). In the region of Beauceville, 45 to 50 miles south-southeast of Quebec, clay at a number of localities is covered by till (MacKay, 1921, p. 54). On the St. Francis River near Pierreville, 50 miles northeast of Montreal, boulder clay overlies glacifluvial deposits (Keele, 1915, p. 81). The till beds indicate oscillations of the ice border and very slow uncovering.

### The Region Between Lake Ontario and the Line Mattawa River-Maniwaki

In the region between Lake Ontario and the Mattawa River-Maniwaki the rate of the ice recession and the actual length of the uncovering are not known because of scarcity of varved clay and because of halts and readvances marked by moraines, overridden clays, and extensive wash plains. Lack of agreement between very long varve series observed at places close by each other makes it clear that the uncovering of this belt took a long time. The

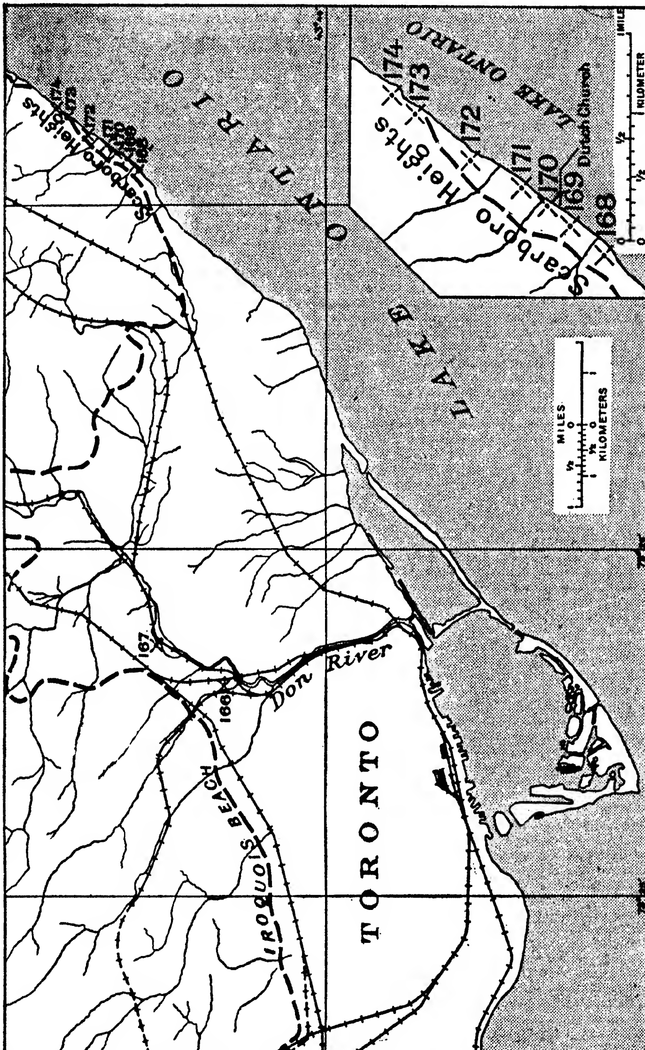



FIG. 17—Position of the localities examined in Toronto, Ont. (localities 166–174). (Base map and the Iroquois Beach from Coleman, 1913a.)

several and in part prolonged stages undergone by the Great Lakes during the corresponding time indicate the same thing. A good idea of the length of this period is obtained from the estimates of the cutting of the Niagara Gorge, whose different sections correspond to certain stages in the history of the Great Lakes, stages in part determined by the positions of the ice border. (Fig. 4, p. 99; Fig. 29, p. 164).

The glacial beds at Toronto, localities 166–174, overlying the interglacial Don and Scarboro beds,[1] are of as yet undetermined age. Since they are glacial land and fresh-water deposits they contain no fossils. Except for some sand beds the deposits under discussion are all through the series hard-packed and have undergone partial consolidation and form magnificent, precipitous bluffs, where attacked by Lake Ontario. The beds can be worked with sharp spade only with great difficulty. The uppermost beds do not differ in any respect from the lowermost, but all appear to date from the same glaciation; the alternation of stratified deposits and till may mark only slight oscillations of the ice border. However, nothing is known to contradict the possibility that the beds date from two or three different glacial epochs. Overriding of varved clay by advancing ice border does not compress it perceptibly. Compared with the normally soft and mellow late-glacial clays, silts, and boulder clays those under discussion give the impression of very considerable age. The whole series of beds would seem to be of pre-Wisconsin age, most probably derived from the third or next to the last, the Illinoian-Iowan, glaciation. As pointed out by W. A. Johnston, however, the advanced consolidation might be due to exceptionally favorable conditions for this process, drainage, texture of the material, presence of lime, etc., rather than to long time. Complete ab-

[1] On the basis of their rich fauna and flora, the Don and Scarboro beds have been attributed to the third, or Sangamon, interglacial epoch—the interval between the Illinoian and the Iowan glaciations (Leverett, 1903; F. C. Baker, 1920, p. 327; Hay, 1923, p. 283; Coleman, 1926, p. 26; cf. Coleman 1913a; 1915, p. 252). However, if, as Leverett (1926, p. 115) is now inclined to think, the Illinoian drift and the Iowan drift date from the same glaciation, the interglacial Toronto beds may rather be derived from the second and longest interval, the Yarmouth, falling between the Kansan and Illinoian glaciations. The Toronto beds have just been attributed to the Yarmouth also by Coleman (1927, p. 401).

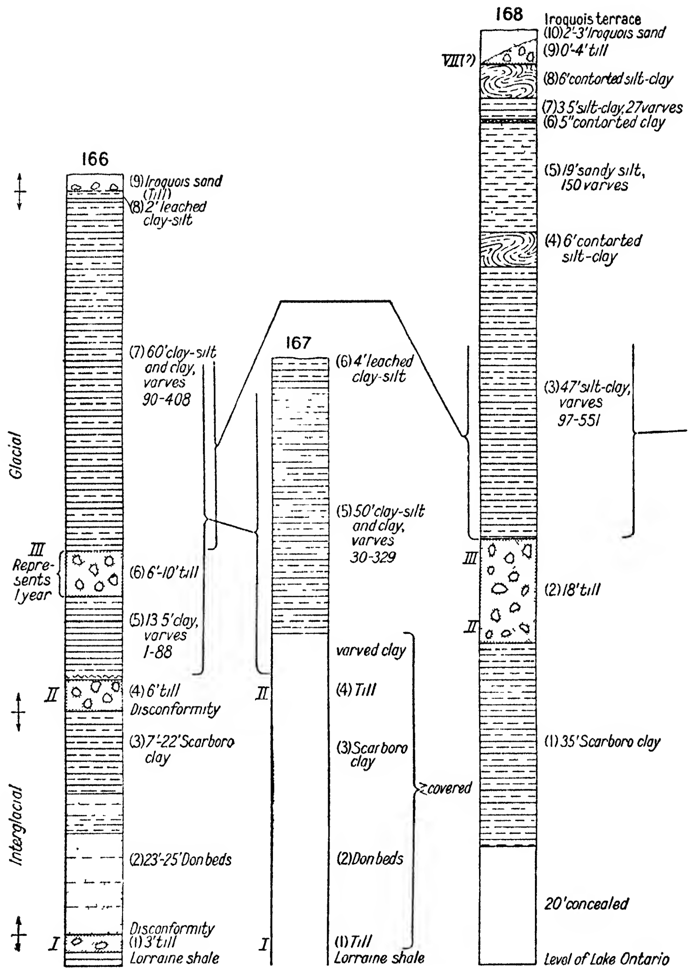



FIG. 18

FIGS. 18–20—Profiles of the interglacial and glacial beds at Toronto, Ont. (localities 166–174).

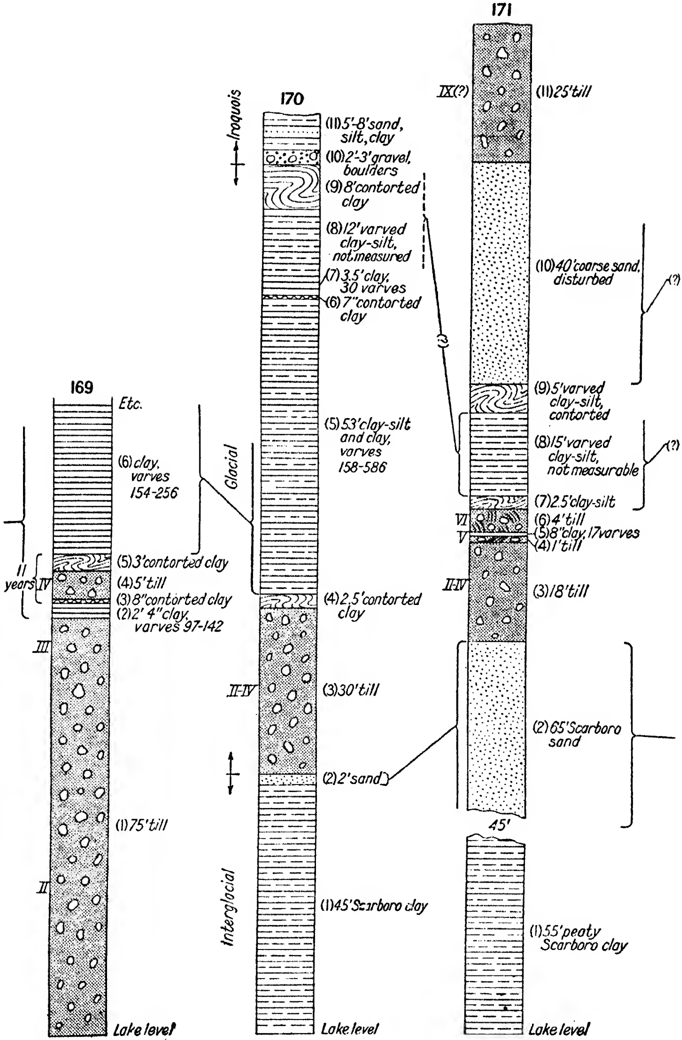
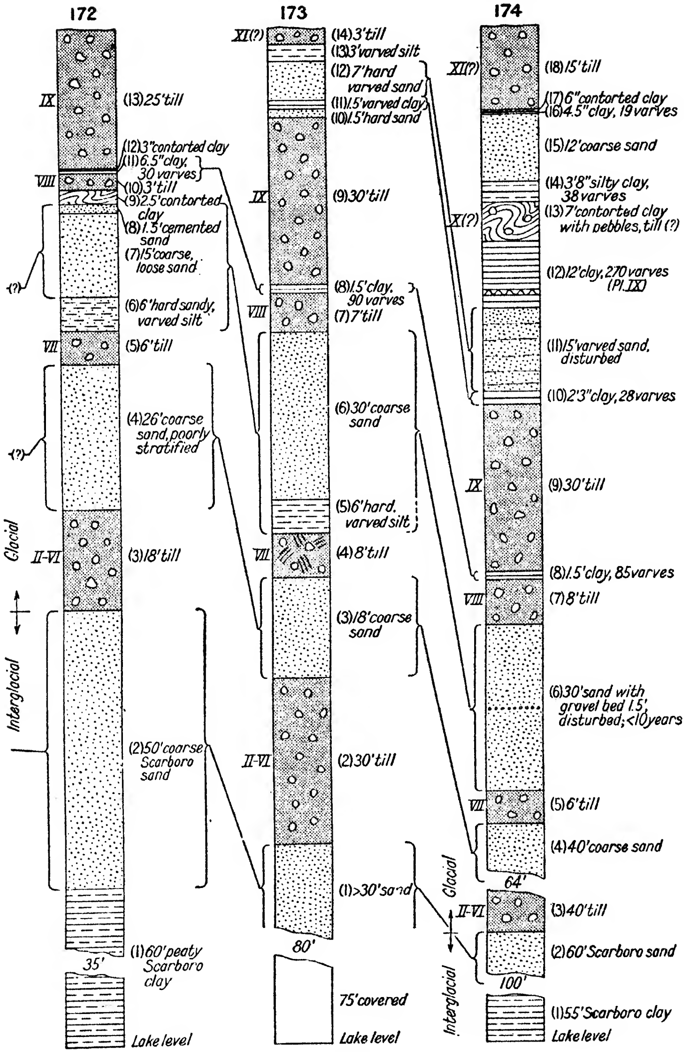

169
Etc.
(6) clay, varves 154-256
11 years
IV
(5) 3' contorted clay
(4) 5' till
(3) 8" contorted clay
(2) 2' 4" clay, varves 97-142
III
(1) 75' till
II
Lake level
170
Iroquois
(11) 5'-8' sand, silt, clay
(10) 2'-3' gravel, boulders
(9) 8' contorted clay
(8) 12' varved clay-silt, not measured
(7) 3.5' clay, 30 varves
(6) 7" contorted clay
(5) 53' clay-silt and clay, varves 158-586
Glacial
(4) 2.5' contorted clay
II-IV
(3) 30' till
(2) 2' sand
Interglacial
(1) 45' Scarboro clay
Lake level
171
IX(?)
(11) 25' till
(10) 40' coarse sand, disturbed
(?)
(9) 5' varved clay-silt, contorted
(8) 15' varved clay-silt, not measurable
(?)
(7) 2.5' clay-silt
VI
(6) 4' till
V
(5) 8" clay, 17 varves
(4) 1' till
II-IV
(3) 18' till
(2) 65' Scarboro sand
45'
(1) 55' peaty Scarboro clay
Lake level

FIG. 19

sence of glacial Wisconsin deposits in the region, as the other view requires, would surely be strange. Nowhere in Canada has indisputable record of more than two Quaternary glaciations been found, and the Don beds are underlain by a Quaternary till. Therefore the whole set of beds above the interglacial ones were perhaps deposited during oscillations of the waning last ice sheet.

The beds and the varve correlations show that, when the ice still covered the region just north of Toronto and filled the greater part of the Ontario basin damming up its water, a number of oscillations of the ice border took place, since, as has been stated, all the glacial beds probably belong to the same glacial epoch. Whether locality 166 or 167 was the first one to be uncovered is not known, since at locality 166 the base of the clay has suffered through sliding and since at locality 167 the clay measurement begins about three feet above the till, the lower part of the clay being covered. For although the varve series from section 166 presented in Plate IX rests directly on till, a lower, partly disturbed series of varves was measured above till bed II 50 yards from the profile. More than 90 years after the first uncovering of these localities the ice for one year expanded from the south over section 166, but not over section 167, depositing till bed III (Fig. 18; Pl. IX, curves 1). Eight years later the localities 168 and 169 were first freed from the ice. About 50 years after this event the ice border readvanced over section 169 but not over locality 168. The oscillation was of very brief duration, perhaps only of one or a few years, for the till deposited, bed IV, and the contorted clays below and above together represent only 11 years. The fluctuation was limited to the zone between profiles 168 and 170, that is amounted to a few hundred yards. The time of the first uncovering of section 170 is not exactly known, since the lowermost two-and-a-half feet of the clay are contorted. The first measurable varve is number 158. At localities 166–170 deposition of varved clay then proceeded without interruption for a considerable time, since at section 168 about 640 varves have been measured. During this long time little or no clay was deposited at localities 171–174, which appear to have been largely



172
IX
(13)25'till
(12)3"contorted clay
(11)6.5"clay, 30 varves
VIII
(10)3'till
(9)2.5'contorted clay
(8)1.5'cemented sand
(7)15'coarse, loose sand
(?)
(6)6'hard sandy, varved silt
VII
(5)6'till
(4)26'coarse sand, poorly stratified
(?)
Glacial
II-VI
(3)18'till
Interglacial
(2)50'coarse Scarboro sand
(1)60'peaty Scarboro clay
35'
Lake level
173
XI(?)
(14)3'till
(13)3'varved silt
(12)7'hard varved sand
(11)1.5'varved clay
(10)1.5'hard sand
IX
(9)30'till
(8)1.5'clay, 90 varves
VIII
(7)7'till
(6)30'coarse sand
(5)6'hard, varved silt
VII
(4)8'till
(3)18'coarse sand
II-VI
(2)30'till
(1)>30'sand
80'
75'covered
Lake level
174
XII(?)
(18)15'till
(17)6"contorted clay
(16)4.5"clay, 19 varves
(15)12'coarse sand
(14)3'8"silty clay, 38 varves
X(?)
(13)7'contorted clay with pebbles, till (?)
(12)12'clay, 270 varves (Pl. IX)
(11)15'varved sand, disturbed
(10)2'3"clay, 28 varves
IX
(9)30'till
(8)1.5'clay, 85 varves
VIII
(7)8'till
(6)30'sand with gravel bed 1.5', disturbed; <10 years
VII
(5)6'till
(4)40'coarse sand
64'
Glacial
II-VI
(3)40'till
Interglacial
(2)60'Scarboro sand
100'
(1)55'Scarboro clay
Lake level


FIG. 20

covered by the ice. Locality 171 probably was uncovered not very long after section 170, but the ice edge soon readvanced, crumpling up the clay and thus forming till bed V, which contains lumps of varved clay. Before reaching locality 170, however, the ice border retired anew, and a bed of clay was laid down. This probably happened before the deposition of varve 300. Later, during a new readvance of the ice front, recorded by till bed VI, the clay was partly torn up. Localities 172, 173, and 174 seem to have been covered by the ice up to this time and to have been freed during the subsequent ice retreat. After the deposition of a bed of coarse sand, which need not have taken many years, and of some varved clay, as suggested by the clay lumps in till bed VII, profile 173, the ice border readvanced over sections 172, 173, and 174 and probably over 168, 169, and 170, depositing till bed VII. Then the ice front again retired. In the sediments now formed the sand bed 6, profile 174, records less than ten years, while beds 6 to 9, profile 172, may represent a few decades. So a new readvance set in, and till bed VIII was formed. Again the ice edge withdrew. It left sections 172 and 173 the same year and locality 174 at practically the same time. The ice melting now was very small, the average thickness of the varves being only one-fifth of an inch (5 mm.). After 90 years or more the ice front pushed forward, as till bed IX testifies. This time the ice probably expanded over section 171 and perhaps still farther westward. During the subsequent withdrawal of the ice clay was first deposited for about 30 years, then sand for a few years, and again clay for some 270 years (beds 10, 11, and 12, profile 174). The sand indicates approach of the ice front or movement of the mouth of the glacial river. Bed 13, profile 174, may mark readvance of the ice edge or the grounding of an iceberg. Subsequently, at section 174, clay was laid down for 38 years, coarse sand for a few years, and again clay for 19 + x years. Finally, the ice pushed forward for the last time, so far as these records indicate, and the capping till in profile 174 and probably also that in section 173 were formed.

Boulders in the Iroquois sand at locality 166 may represent

residuals of a till bed, and thus indicate that the ice advanced over the place more than 408 years after it left for the first time.

The varved clays that remain in the whole series seem to record about 1200 years. The crumbled clays represent an unknown, but probably not great, number of years, and the till beds another unknown number. Supply and wastage of the ice practically balanced each other. The ice front seems to have lain to the north and the east of the profiles and outside the present coast line and from these positions to have pushed forward from time to time. In the cases where the time represented by the till beds has been determined it has been found to amount only to a few years. Therefore it is believed that the intercalated till beds correspond to much shorter time than do the varved clays and that all the glacial deposits with the exclusion of the first till layer on the Scarboro bed in each section, in other words the seven or eight oscillations, represent about 1500 years.

The best method for obtaining an idea of the rate of the uncovering of southern Ontario is, as stated, to correlate positions of the ice border with stages of the upper Great Lakes and to apply the time estimates based on the cutting of the Niagara Gorge. On an earlier occasion attempts have been made to reconstruct the position of the ice border at a few important stages of the lakes (Antevs, 1925b, Fig. 27, p. 74). These positions can now be somewhat better defined. At the inauguration of the Kirkfield stage of Lake Algonquin, which was brought about by the opening of the Trent Valley, the ice front had reached the central part of Stony Lake, 88 miles northeast of Toronto. Judging from the direction of the striae it trended northwest-southeast. It was tentatively drawn north of locality 33, Espanola, on the north side of Lake Huron. However, as records of upheaval of land during the Champlain stage occur as far west as Pembroke, the lack of indications of oscillations of land in the thick deposit at Espanola (Antevs, 1925b, pp. 104, 123) perhaps makes it more probable that the region was first uncovered after the end of the first marine deep-water stage, subsequent to the end of the Kirkfield stage. (Fig. 4, p. 99; Fig. 29, p. 164).

During the later part of the life of Lake Iroquois the ice retreat appears to have been insignificant, for the Iroquois beach does not extend northeast of Havelock (Coleman, 1904, p. 352), and a few morainic lines occur above the Iroquois shore north of the Adirondacks (F. B. Taylor, 1924, p. 646). Expansion of an

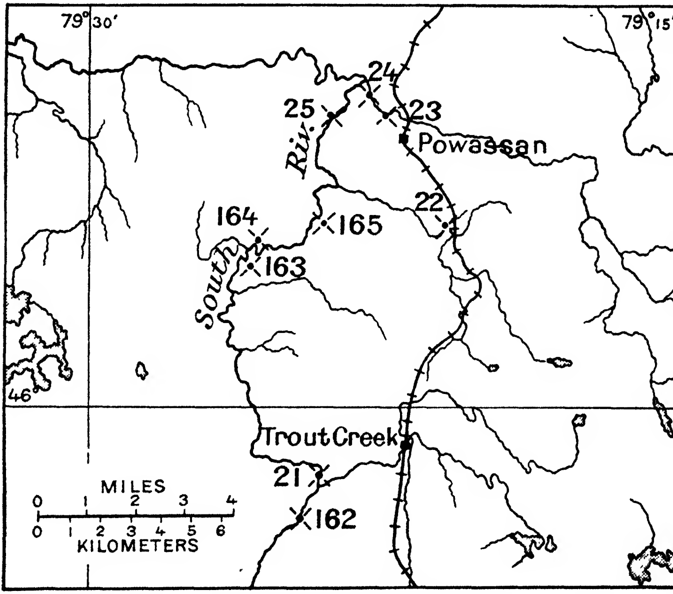



FIG. 21—Position of the localities examined on the South River south of Lake Nipissing, Ont. (localities 21–25, Antevs 1925b, and 162–165).

ice lobe in the St. Lawrence Valley just northeast of Lake Ontario to such a degree as is suggested by F. B. Taylor (1924, Pl. 1, p. 644) does not seem likely, judging from the direction of the striae (see Antevs, 1925b, Fig. 27, p. 74) and from the morainic lines. Six miles north of Kingston there is on the highway a moraine trending slightly north of east. It consists largely of outwash coming from ice lying on the northern side. Thirteen miles north-northeast of Kingston and two miles south-southwest of Battersea

another moraine is to be found running somewhat south of east. Seventeen miles north of Prescott, four miles southeast of Kempton, and five miles northeast of Kempton there are morainic lines trending north of east to nearly northeast and southwest.

The position of the ice front at the end of Lake Frontenac or the beginning of the marine stage in the St. Lawrence lowland is particularly important. The ice edge appears to have stood somewhere south of Delson, eight miles south of Montreal. Here ten feet of marine clay with numerous shells of *Portlandia arctica* are exposed in a clay pit, while till outcrops close by in the bottom of the river at the same level as the bottom of the pit. The lowest varves at localities 151 and 152, 44 miles west of Montreal were undoubtedly laid down in perfectly fresh water. The water certainly was not brackish yet some 125 years after the uncovering, though it was possibly growing so; for, although this part of the clay is indistinctly varved, this circumstance is perhaps due to the thinness of the varves and the fineness of the material. Accordingly, and since the clay would have become massive a little above the bottom, especially as the localities lie in the central part of the Champlain Sea area, the uncovering may have occurred before the breaking in of the sea. On the other hand, the ice border does not seem to have reached to two miles north of Hawkesbury, for the clay resting on till on Kingham River and the highway appears to have been laid down in brackish water, though no shells were observed. In the Ottawa region the ice may have reached just beyond the Ottawa River, as fresh-water clays occur beneath the marine clays south of the river but not five miles north of it (Antevs, 1925b, p. 64). Farther west the ice border may have stood south of Quyon and Renfrew at the critical time. At Quyon shelly clay appears to rest immediately on bed rock (cf. p. 100). At the brickyard in the eastern edge of Renfrew the clay directly on the till shows faint varvity and on the whole is almost massive, indicating deposition in brackish water, though no shells were observed. The absence of distinct varves at the base is noteworthy because of the situation so far up the estuary.

During the first marine deep-water stage the ice border withdrew to beyond Brennan station, 28 miles north-northwest of Ottawa, and to beyond Pembroke, where deposits of the lower marine clay were observed. How far north of these places it retired is not known from direct observation, and the conditions as known at present are open to two explanations. The vast wash plain between Forester Falls and Beachburg, 15 miles east-southeast of Pembroke, as shown by the level surface built up practically to the water surface of the first marine deep-water stage, reaches about 560 feet (171 m.) altitude. Glaciaqueous deposits occurring north of here reach approximately the same level and thus were perhaps formed during the same stage. Among these is an extensive sedimentation plain at about 550 feet (168 m.) of altitude a few miles northeast of Fort Coulonge on the north side of the Ottawa River. Others are wash plains just west of Pembroke, and from six miles southeast of Petawawa to seven miles northwest of this place, rising to between 470 and 500 feet (143 and 152 m.). Sand plains extend on the north side of the Ottawa River eastward all the way to Kazubazua, where the surface attains an elevation of 600 feet (183 m.) (Ells, 1907, p. 37).

In the Gatineau Valley glaciaqueous deposits extend up to the vicinity of Gracefield. North of here such are lacking along the railway, though the altitudes of the Blue Sea Lake, Burbridge, and Farley stations are 551, 571, and 573 feet (168, 174 and 175 m.), respectively, while that of Kazubazua lying on the wash plain is 600. The altitude to which the deposits occur therefore decreases north of Kazubazua. At Maniwaki, at the end of the railway 70 miles north of Ottawa, glacial sediments occur up to about 580 feet (177 m.) above sea level, and the water body in which they were laid down may have reached at least the altitude of 600 feet (Wilson, 1924, p. 128). These conditions suggest a new rise of the water level. The sediments at Maniwaki may accordingly date from the second marine deep-water stage. The clay at Maniwaki was surely formed in perfectly fresh water, but it cannot be concluded from this fact that it was laid down in a

separate lake, since a bay of the Champlain Sea so far inland would no doubt have had fresh water. During the marine shallow-water stage the ice border may have retired between

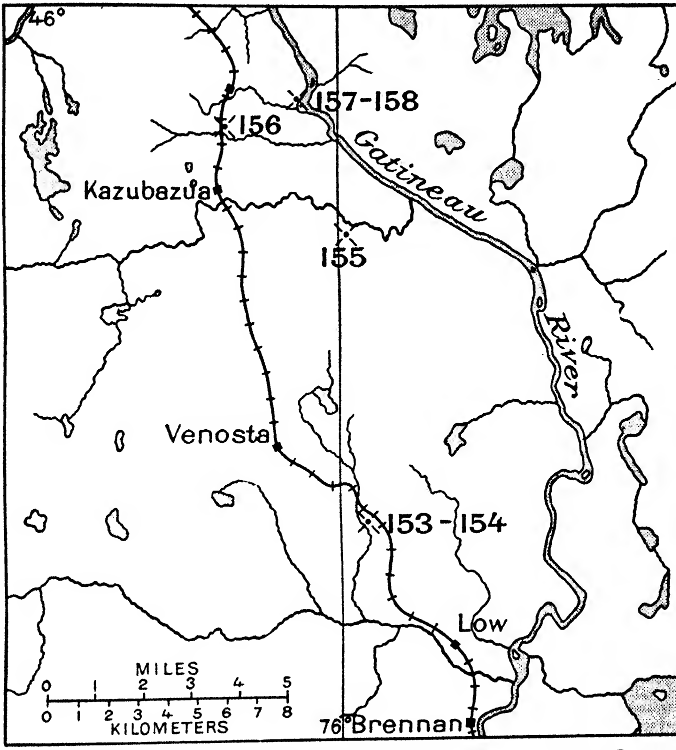



FIG. 22—Position of the localities examined in the Kazubazua region, Que., 40 miles north of Ottawa (localities 153–158).

Gracefield and Maniwaki. At the end of the first deep-water stage it may have lain somewhat north of the Kazubazua and Petawawa wash plains and have trended about east and west. This is the one explanation.

The other possible explanation seems equally valid. As the

land was rising during the uncovering from the ice, there probably was a belt in which the limits of the first and the second marine deep-water stages coincided and north of which the latter lay above the former. These conditions were evidently dependent upon the situation of the ice border at the end of the first deep-water stage. Therefore, agreement in height of the upper limits is no sure evidence of approximate contemporaneity. The wash plains at Kazubazua and Petawawa perhaps date from the second marine deep-water stage. Whether glaciaqueous deposits occur on the Gatineau all the way to Maniwaki is not known to the writer. The Champlain Sea perhaps never extended a branch to the Maniwaki region. Maniwaki lies in a flat basin. Lac des Oblats and the Cedar Lakes southwest of Maniwaki drain northward. In late-glacial time the earth's crust inclined northward more than now; and since then the Gatineau no doubt has cut down its channel. These conditions make it possible that a lake extended over the lower portions of the Maniwaki region subsequent to the Champlain stage. According to this view the ice edge may have retired between Gracefield and Maniwaki as the uplift that ended the Champlain stage was in process.

The ice recession during the first marine deep-water stage was broken by halts, as indicated by moraines and wash plains some of which have already been alluded to. North of St. Thérèse, 18 miles northwest of Montreal, a probable moraine consisting largely of outwash was observed. To the north of it, between St. Thérèse and St. Janvier there is a sand plain. Eighteen miles northeast of Ottawa and three miles northwest of Buckingham, on the west side of Lièvre River, a deposit of outwash one-half mile wide with a marked ice contact on the north side is to be found. Two miles farther north on the east side of the river there is a similar deposit. The ice contact here forms a bluff 120 feet high. Fifty miles west-northwest of Ottawa, on the north side of the Ottawa River in the southern part of the townships of Clarendon and Bristol, large plains of sand are reported (Ells, 1907, p. 38). To the southwest of this point, at the northwestern edge of Renfrew, there is a deposit of outwash with large kettles.

The enormous gravel plain between Forester Falls and Beachburg, already mentioned, has a distinct ice contact with numerous kettles, among which are the basins of the lakes indicated on the maps a little southeast of Beachburg. To the west and southwest outwash deposits have extensive distribution on the western side of Muskrat Lake and around Golden and Round Lakes (Ells, 1907, p. 38). The wash plains extending from Petawawa to Kazubazua, the largest of them all, indicate a marked halt in the ice recession. Whether these latter date from the first or the second marine deep-water stage is, as discussed, unknown. If they belong to the second, little retreat may have taken place during the shallow-water stage.

The rate of the ice retreat north of the large wash plains is suggested by a few conditions. At and north of Burbridge, nine miles south of Maniwaki, there is much sand and gravel, Clement Lake being a glacial pond. On the southern side of the Ottawa River northwest of Petawawa outwash deposits were observed from the train at Chalk River, a mile west of Bass Lake, a mile and a half south of Ashport, just east of Mackay, on the southern side of Brennan Lake, just east of Bissett, and at Deux Rivières. It is not known, however, whether they all signify halts in the retreat. During the latter part of the second marine deep-water stage the ice border was practically stationary for a long time, for, although the Algonquin shore line is recorded at a high level five miles northeast of North Bay, the Mattawa Valley did not become ice-free until after the great upheaval that put an end to Lake Algonquin had been nearly completed. The thinness of the varves at locality 159, four miles north of Maniwaki, may possibly be due to trapping of part of the glacier mud in a basin farther north or to another course of the material; but more probably it means that the ice melting was slight. The remarkable evenness of the clay both as regards texture of the material and thickness of the varves shows that it must have been transported by a land river discharging continuously at about the same place.

As has been pointed out on an earlier occasion (Antevs, 1925b, p. 78), the uplift of the land that divided the marine stage into

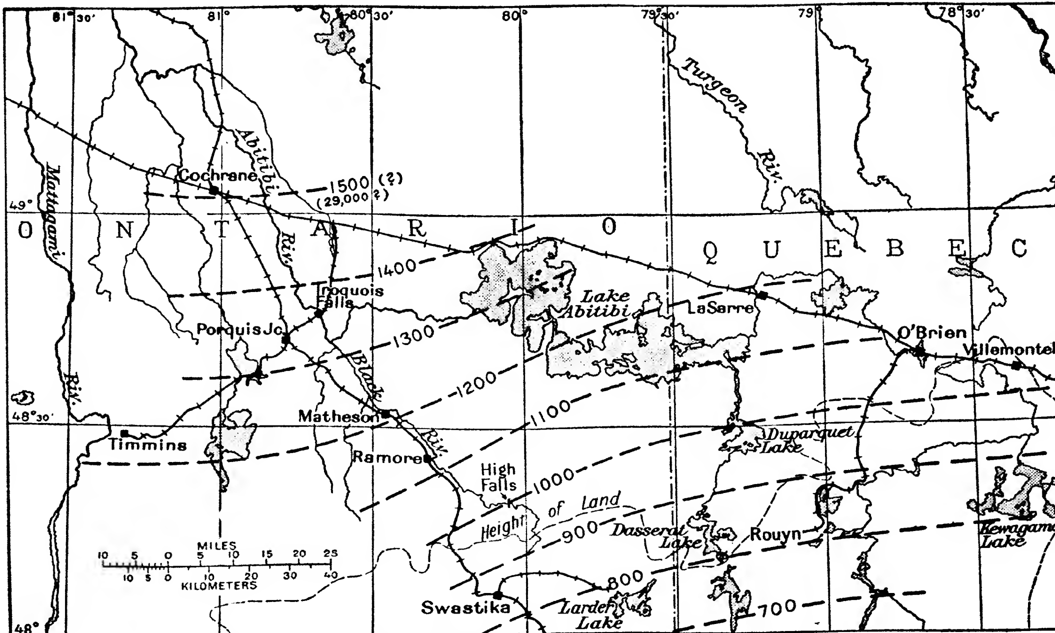



FIG. 23—Map of the Abitibi region, Ontario and Quebec, showing position of the last retreating ice border every 100 years. Year 1 is the year of uncovering of the mouth of the Montreal River in Lake Timiskaming, 47° 8′ N., 79° 27′ W. (locality 63). Continuation southward, etc., in Figs. 36 and 37, pp. 106 and 116, Antevs, 1925b.

three substages may have been the same as that which caused the outlet of Lake Algonquin to change from the Trent Valley to Port Huron. The ice retreat from Stony Lake, 88 miles northeast of Toronto, to the region of Kazubazua or to that south of Maniwaki may have corresponded to the Kirkfield stage of Lake Algonquin and to the Old Narrow Gorge of the Niagara River and may have taken about 10,000 years if the whole gorge represents 25,000 years (F. B. Taylor, 1913, p. 53; 1913b, p. 24). The retreat to beyond the Mattawa Valley and to the region south of Maniwaki or north of Maniwaki corresponds to the later part of the marine shallow-water stage and the second marine deep-water stage, i. e. to the Port Huron stage of Lake Algonquin and to the Lower Great Gorge of the Niagara River. It may thus represent 2500 to 3000 years (F. B. Taylor, 1913b, pp. 24, 25).

### Northern Ontario and Quebec

The rate of the ice recession increased in the Timiskaming region, being 454 feet (138 m.) a year between locality 63 in the southern part of Lake Timiskaming and locality 82, Englehart, and 542 feet (165 m.) annually between the later latter locality and locality 118, La Sarre, northeast of Lake Abitibi. It was slower in the shallow-water and in the supra-aquatic region along the Temiskaming and Northern Ontario Railway than in the deeper water in westernmost Quebec. Between Englehart and the divide the rate may have been rather uniform judging from the evenness of the varve curves. (Fig. 3, p. 98).

North of the height of land the rate increased somewhat with the growing depth of water, amounting between locality 95, Matheson, and locality 134, Frederick House River, to about 645 feet (197 m.) annually and between this latter locality and number 136 to some 820 feet (250 m.) annually, provided locality 136 is correctly connected, as is probable, and provided this locality became ice-free about year 1500. If this be so, the retreat may to a large extent have taken place by calving, for the decreasing thickness of the varves 1450–1527 suggests diminishing melting. This latter may have been abruptly followed by rapid ice melting,

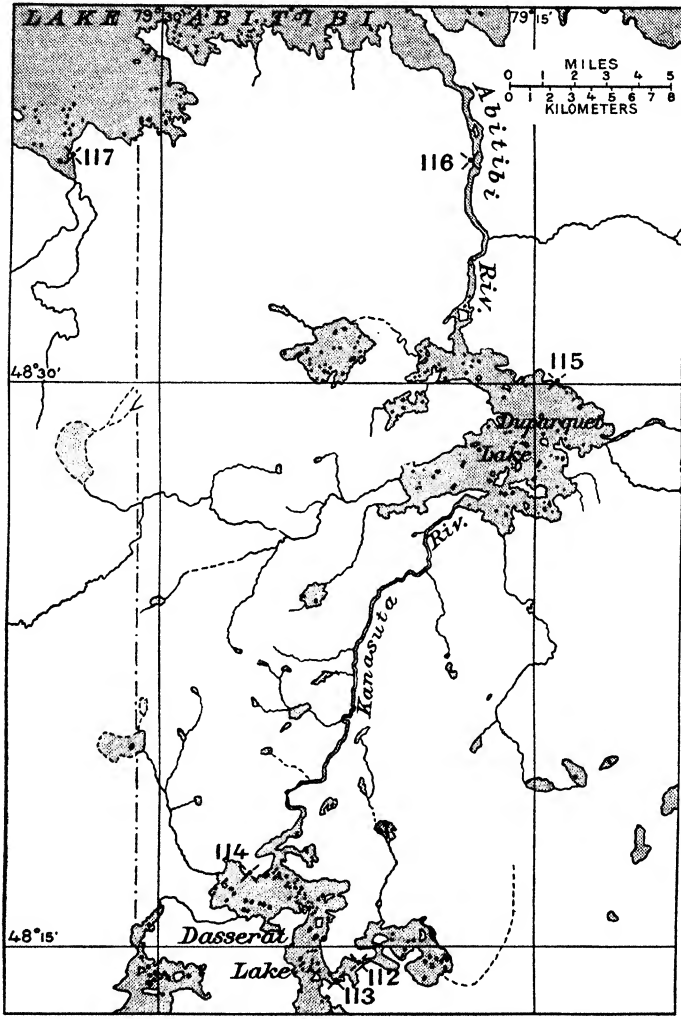



FIG. 24—Position of the localities examined in the region south of Lake Abitibi, Que. (localities 112–117).

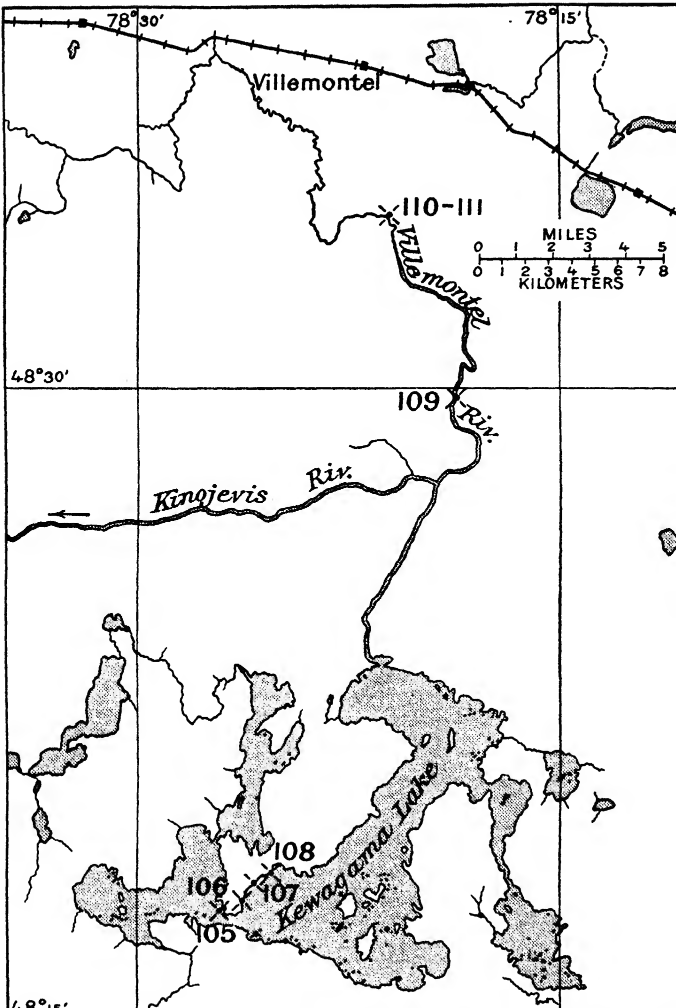



FIG. 25—Position of the localities examined on Kewagama Lake and Villemontel River, Que. (localities 105–111).

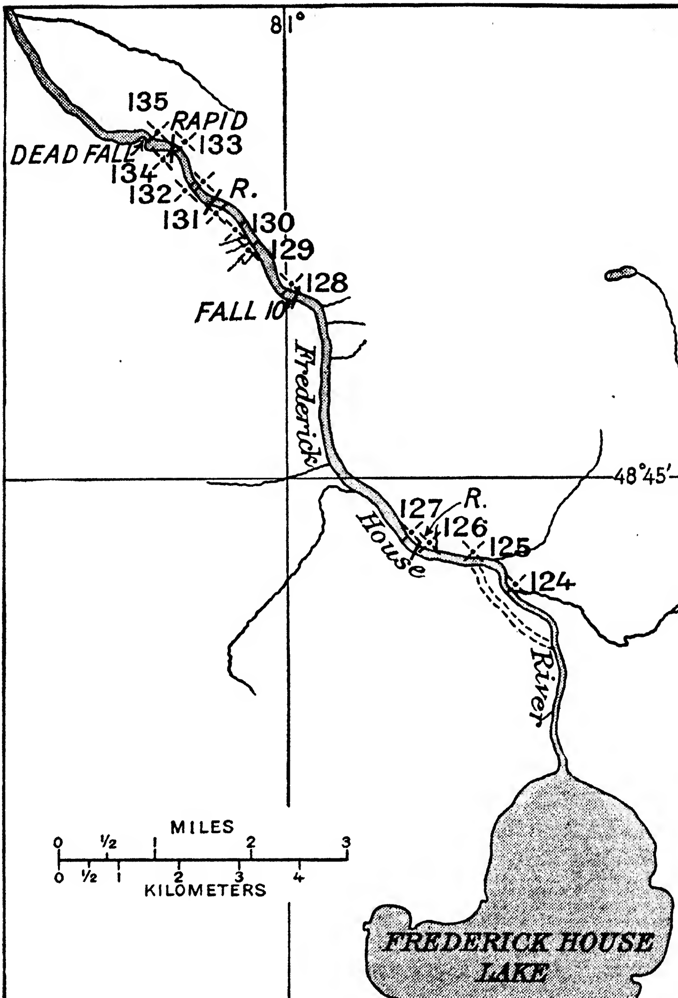



FIG. 26—Position of the localities examined on the Frederick House River, Ont., northwest of Porquis Junction (localities 124–135). Measurements 128–135 were obtained in the deep gorge formed after the diversion of the river in 1909 (p. 171).

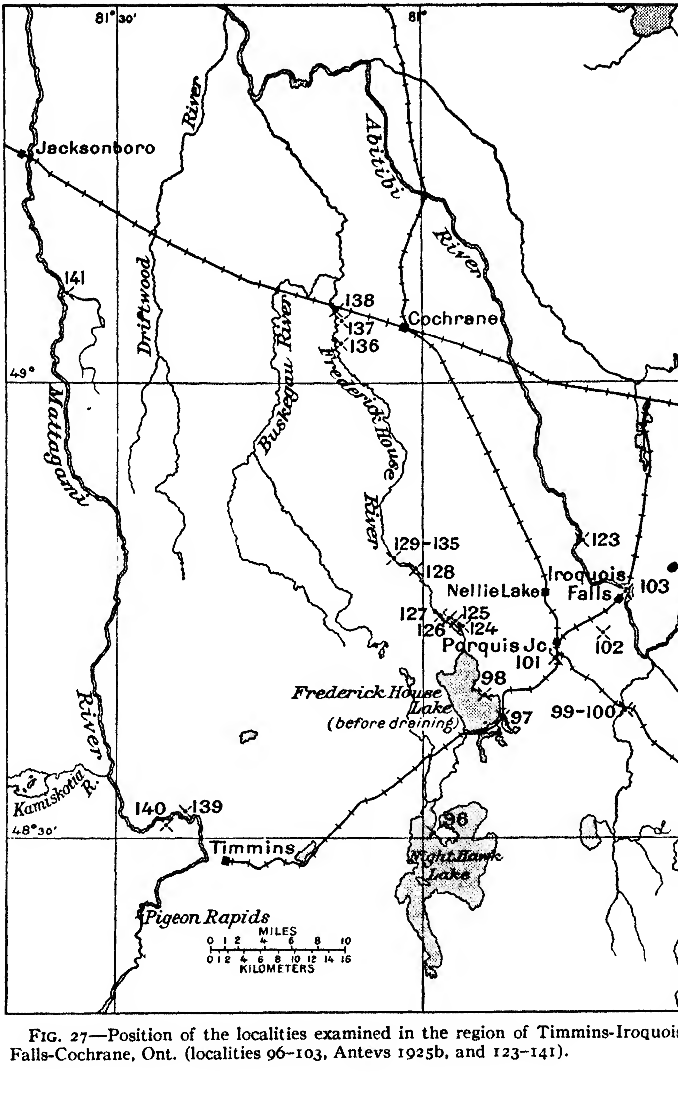



FIG. 27—Position of the localities examined in the region of Timmins-Iroquois Falls-Cochrane, Ont. (localities 96–103, Antevs 1925b, and 123–141).

RATE OF RETREAT OF THE ICE BORDER

| Localities | Number of Bottom Varves | Distance | Time of Retreat in Years | Rate of Retreat a Year |
|---|---|---|---|---|
| 63 to 82 | 1, 594 | 51 miles = 269,280 feet (82.1 km.) | 593 | 454 feet (138 m.) |
| 82 to 118 | 594, 1208 | 63 miles = 332,640 feet (101.4 km.) | 614 | 542 feet (165 m.) |
| 63 to 118 | 1, 1208 | 118 miles = 623,040 feet (189.9 km.) | 1208 | 515 feet (157 m.) |
| 82 to 95 | 594, 1212(?) | 58 miles = 306,240 feet (93.3 km.) | 618 | 496 feet (151 m.) |
| 95 to 134 | 1212(?), about 1392 | 22 miles = 116,160 feet (35.4 km.) | About 180 | 645 feet (197 m.) |
| 134 to 136 | About 1392, about 1500 | 17 miles = 87,760 feet (27.4 km.) | About 110 | 820 feet (250 m.) |

for the extraordinary increase in the thickness of the varves beginning with varve 1528 seems to be chiefly due to this, as it is equally marked in independently fed parts of the ancient lake, and the character of the clay does not undergo any change. The varves remain thick for about 200 years. During this time of apparently rapid retreat the ice border may have retired far north of Cochrane, though locality 137, west of Cochrane, is the northernmost place where varved clay correlated with the Timiskaming series has been observed.

At about the year 1700 the varves begin to grow thin either because of diminished ice melting or because of great distance from the ice front. Which was the case cannot be determined with certainty. On the one hand, the ice front subsequently advanced to 23 miles south of Cochrane, to locality 127, which was reached after year 2027. On the other hand, Lake Barlow-Ojibway, or Lake Ojibway, about the year 2025 was drained

across the ice to James Bay (cf. p. 103). As this drainage took place perhaps when the ice front stood far north, rather than during or after its renewed advance it is probable that advance did not set in until after year 2027. However, it is possible that recession had changed into advance as early as about year 1750. If so, the forward motion may have been halted for a short time about years 1840 and 1850, as considerable melting then took place.

The ice expanded on the Frederick House River to locality 127 but not to locality 125. It probably also reached to ten miles northwest of Timmins, where on the Mattagami River two miles south of the mouth of the Kamiskotia River a slide presents varved clay with cobbles kneaded into it. It did not reach the localities 139 and 140 a few miles to the southeast. On the Abitibi River the ice front advanced to a point between one and two miles south of Iroquois Falls, for the clay is strongly contorted a mile south of the town but undisturbed two miles south of it. On the main railway line it reached to two and a half miles north of Porquis Junction, to the southern edge of the sand and gravel deposits at Nellie Lake (see Knight, etc., 1919, p. 39, map). The enormous quantity of outwash material here indicates a long halt before retreat started in earnest.

Farther east the ice may have expanded to the belt of wash plains and kames that is reported by Tanton (1919, p. 50) to extend eastward from the northern side of Lake Abitibi, from Desmeloises to Béarn township.

From what has been said it follows that the time occupied by the larger oscillation cannot be determined. But, as the first uncovering of locality 127 took place about year 1360 and the ice edge returned first after 2027, it surely represents more than 670 years and probably much more. The rate of expansion of an ice body is almost wholly unknown, but in this case it may be safe to assume that the rate of advance about equaled that of the earlier retreat.

The large deposits of outwash at the limit of the expansion, as stated, indicated that a long time, probably a few hundred years, passed before recession began again. Of the first stage of this

retreat no varved sediments have been observed. However, profile 138, a few miles west of Cochrane, may date from this retreat. This is practically certain, as the long, good, and characteristic varve curve obtained does not agree in the least with the Timiskaming series, whereas normally the measurements of this series, even when obtained in widely separated regions, show a remarkable correspondence with each other. The clay deposit may have been formed in a small lake which did not extend over the slightly higher land to the south, where this series is lacking. Relatively small area of sedimentation, therefore, may be partly responsible for the great thickness of the varves, though the annual ice melting seems to have been fairly great. However, the retreat did not last long after the uncovering of locality 138, for the varved clay is overlain by boulder clay marking overriding by the ice. The increase of the thickness of the varves beginning year 238 may be due to the approach of the ice front. When the ice front halted and readvanced it may have stood some ten miles north of Cochrane. The oscillation, counted from locality 138, represented more than 260 years. The ice border may have advanced to Cochrane, which is built on a moraine, and to a few miles south of locality 138, where on the road leading west on the west side of Frederick House River and to the east of Buskegau River hummocky morainal deposits with kettles are to be found.

Again the ice front receded, but from this retreat only very thin and small deposits of varved clay have been observed in the Cochrane region (see Antevs, 1925b, pp. 58, 77), so that the rate of the retreat is unknown. However, at locality 143, at Mattice on the Missinaibi River, varved clay much younger than that at locality 138 is overlain by till deposited during a readvance of the ice more than 50 years after the first uncovering. The extent of the oscillation is unknown. Judging from the direction of the striae it may be the same oscillation that is recorded at latitude 50° N. and longitude 79° to 79½° W. by a marked morainal belt which has forced the Turgeon River to turn straight eastward (Tanton, 1919, p. 49). The rate of the subsequent recession, registered by the clay at locality 142 Mattice, is unknown.

# CHAPTER VIII

# THE PROBABLE CORRELATION BETWEEN THE LAST ICE RETREAT IN NORTH AMERICA AND IN EUROPE

The most promising approach to a correlation of the recession of the last ice sheets in America and Europe seems to lie neither in the direct matching of varved clay graphs from the two continents—because yearly temperature variations can hardly be expected to have measurable agreement on opposite sides of the Atlantic—nor in the mere comparison of the major moraines of recession as regards number and spacing; but rather in a critical combination of these two methods of study. If we employ De Geer's method of varve measurements and graphs in climatically uniform areas, to work out partial chronologies of the more exact sort wherever possible, and then fill in the inevitable gaps in our lines by estimates based upon other kinds of evidence, such as the Niagara Gorge, we get separate chronologies and climatic records for America and for Europe. By comparing these, but not overlooking related glaciological studies, we may find out whether the advances, halts, and retreats of the receding ice sheets were indeed synchronous. This chapter undertakes to make such a correlation.

## Summary of the Last Ice Retreat in North America

Owing to lack of clay deposits and to absence of exposures the length of time represented by the terminal moraines in eastern North America and by the recession of the ice border from these to Hartford, Conn., is not directly measured. From comparison with other moraines whose time factor is known it appears likely that each of the two moraines in Long Island and the intervening retreat represents several hundred years, and the whole morainic belt about 2000 years. (Fig. 5, p. 108; Fig. 29, p. 164.)

Between Hackensack, N. J., and Haverstraw, N. Y., the annual

retreat averaged at most 62 feet (19 m.). South of Hackensack it may have been still less. Between Haverstraw and Newburg it amounted to less than 134 feet (41 m.). The withdrawal was interrupted at times, as especially observed at Middletown, Conn., in Rhode Island, and in southeastern Massachusetts. The retreat from the terminal moraines to Newburg and Hartford may have taken about 5500 years.

From Hartford, Conn., the rate of the retreat is determined, except for a gap of a few hundred years on the line Claremont-Lake Winnepesaukee, all the way to St. Johnsbury, Vt., a distance of 185 miles (298 km.) (Antevs, 1922, pp. 74–84). The main line of measurements follows the Connecticut River, but long parallel series were obtained in the Hudson and the Merrimac valleys. The time occupied by this recession was about 4100 years. The average rate of the melting back was 238 feet (73 m.) annually, or 22 years to a mile. The actual yearly rate varied from less than zero to 1100 feet (335 m.).

Between Hartford and Springfield the annual recession amounted to about 243 feet (74 m.). Then it decreased; and at Northampton, Mass., the ice front halted and moved forward. The oscillation represented about 350 years. The renewed retreat soon became fairly rapid, amounting in northern Massachusetts to 193 feet (59 m.) a year. The rate increased to about 370 feet (113 m.) annually in the Bellows Falls region. Thereafter retardation set in anew. This resulted in halt and advance at Claremont-Lake Winnepesaukee for probably at least 400 years. The subsequent recession during the first 100 or 200 years attained the high annual rate of 615 feet (188 m.). After a slight decrease it increased again, and some 300 years after the end of the halt it reached 1100 feet (335 m.). This is the greatest speed of melting back observed in North America. Then decrease followed, resulting in halt and advance at St. Johnsbury. This constituted the beginning of a number of oscillations of the ice front in a broad belt in northernmost New England and southernmost Quebec and in the region between Lake Ontario and the Mattawa River.

Before going further the peripheral belt in the Great Lakes region should be briefly dealt with. Morainic belts marking successive stationary positions of the ice front are mapped and correlated by Leverett, Taylor, and others in the Great Lakes region from the Dakotas eastward to western New York (Leverett and Taylor, 1915, p. 62, Pl. 5). Among the marked moraines the Port Huron morainic system and its continuation, the Alden moraine, as varve correlations, directions of striae, and morainic fragments show, may correspond to the Finger Lakes moraine, the moraines west of the Catskill Mountains, and to the oscillation of the ice border at Northampton, Mass. (Antevs, 1922, p. 96, Pl. 6). The position of the ice border thus may be traced from Wisconsin all the way to western Massachusetts. The Niagara Falls moraine, formed during the readvance that closed the Mohawk Valley outlet of the Great Lakes which shortly before had become opened, and its correlative farther west, the Bay City moraine, or the Tawas moraine (F. B. Taylor, 1915, p. 398), may correspond to the overridden clays and push moraines at Claremont-Lake Winnepesaukee in New Hampshire (Antevs, 1922, p. 99; Goldthwait, 1925).

While no direct evidence for the correlation of the other morainic lines exists, it is from their positions highly probable that the outermost moraine and the Bloomington morainic system of the Middle West correspond to the two morainic belts in Long Island, the Mississinawa morainic system to one of the moraines in southeastern Massachusetts, and the Defiance moraine to the ice oscillation at Middletown, Conn. (Fig. 29, p. 164.)

In the belt from St. Johnsbury and Lake Ontario northward to the central part of Lake Timiskaming the rate of the ice retreat is not chronologically determined because of scarcity of varved clay, halts, and readvances of the ice edge, indicated by wash plains, moraines, and overridden clays. The first oscillation took place at St. Johnsbury. The complete lack of agreement between long varve series from closely located points also makes it evident that the time occupied by the retreat from Lake Ontario to the northern side of the Mattawa Valley was many

thousand years. So do also the several and in part long-continued stages of the Great Lakes and the St. Lawrence lowland. The best time measure for the uncovering of this belt is the cutting of the Niagara Gorge, whose various sections correspond to known stages of the Great Lakes, stages determined by positions of the ice border. The uncovering of Stony Lake, 88 miles northeast of Toronto, inaugurated the Kirkfield stage of Lake Algonquin and the formation of the corresponding Old Narrow Gorge. The retreat to beyond the Mattawa Valley meant the end of Lake Algonquin, the end of the cutting of the Lower Great Gorge. If the formation of the whole Niagara Gorge took 25,000 years, as is supposed, then 12,500 to 13,000 years come on the two sections referred to here. To this is to be added an unknown figure for the retreat of the ice border from St. Johnsbury to the position at the time of the opening of the Trent Valley which in Vermont may have been about half-way between St. Johnsbury and the Canadian frontier. (Fig. 2, p. 95; Fig. 4, p. 99.)

The rate of withdrawal between the Mattawa Valley and the mouth of the Montreal River in Lake Timiskaming, a rocky and stony wilderness without fine glaciaqueous sediments, is not known. Since the distance is 60 miles, the retreat, if moderate, may have taken about 1000 years.

In the Timiskaming region the recession was moderately rapid, averaging 454 feet (138 m.) annually. The rate increased somewhat. From a little south of the divide up to the Transcontinental Railway at Lake Abitibi it attained 542 feet (165 m.) a year. The retreat from locality 63 on Lake Timiskaming to locality 118, La Sarre, a distance of 118 miles (190 km.) took 1208 years, the average annual rate being 515 feet (157 m.). With growing depth of water north of the divide the rate of melting back increased, reaching perhaps as much as 1000 feet (300 m.) a year. Subsequently it fluctuated, as is suggested by variations in the thickness of the varves. When the ice border had reached north of Cochrane, i.e. probably after year 2027 or possibly somewhat earlier, it began to advance, finally extending as far as Iroquois Falls and about two miles north of Porquis

Junction. It reached this point certainly more than 670 years and probably some 1300 years after it first left. After remaining stationary for a few hundred years, judging from the quantity of accumulated outwash, the ice border began to recede. It receded until it stood ten miles or so north of Cochrane, when it halted and advanced anew. This time the ice front reached Cochrane. Before long it began to withdraw, but again it halted, as recorded at Mattice 110 miles (175 km.) west-northwest of Cochrane and at a point 110 miles northeast of Cochrane. Regarding the subsequent retreat no data whatever have been obtained. (Fig. 23, p. 142.)

To sum up, the length of time of the recession up to the first advance at Cochrane may have been as follows:

| | | Years |
|---|---|---|
| The terminal moraines.................. | about | 2,000 |
| From the terminal moraines to Hartford, Conn.... .. ... .. ... ......... | about | 5,500 |
| Hartford, Conn., to St. Johnsbury, Vt.... | | 4,100 |
| St. Johnsbury, Vt., to Stony Lake, Ont... | | x |
| Stony Lake, Ont., to Mattawa Valley, Ont.. | about | 13,000 |
| Mattawa, Ont., to mouth of Montreal River on Lake Timiskaming, Ont... ........ | | x |
| Mouth of Montreal River, Ont., to north of Cochrane, Ont...................... | | 2,000 |
| Total.................................. | | 26,600+x |
| | | Probably 28,000 to 29,000 years |

## Summary of the Last Ice Retreat in Europe

### NORTHERN GERMANY

The extreme limit of the last ice sheet in northern Germany is debated. It is believed by different geologists to be marked by various prominent morainic lines, whose different parts, furthermore, have been differently correlated. The three younger sets of moraines, the Brandenburg, the Poznań, and the Pommerania, do not differ materially in age and certainly

belong to the same, the last, glaciation (Woldstedt, 1925a, pp. 177, 180, 181; our Fig. 30). The Brandenburg moraine forms the southern limit of the lake region of northern Germany. The region outside is much more smooth, of very much older topography. Woldstedt leaves the question open whether the Brandenburg moraine forms the limit of the last glaciation or of a second expansion of the last ice sheet, depending upon whether *an interglacial or a very long interstadial epoch fell between the formation of the Flaming moraine and the Brandenburg.* Gripp (1924, p. 217, Pl. 13), however, after a detailed study concludes that the distinct morphological boundary formed by the Brandenburg moraine constitutes the extreme limit to which the last ice sheet extended. The Brandenburg moraine consists partly of bouldery till ridges, which in some places are steep and up to a few hundred feet high and in other places are hardly noticeable, and partly of wash plains several miles wide. No geochronological studies have been carried out in Germany or Poland; but, judging from moraines whose time of formation is known, the one under consideration may represent from a few hundred years to 1000 years at most. The limited amount of till and outwash shows clearly that the time cannot be very long.

After the ice recession had well started it seems to have proceeded for a time without marked interruptions, for there are only scattered moraines between the Brandenburg morainic system and the Poznań (see Wahnschaffe and Schucht, 1921, Pl. 21). The rate, however, may have been slow. The merging of the Poznań moraine with the Brandenburg both in the west and the east, or its position outside the Brandenburg moraine, suggests that the ice border first retired to far beyond the site of the former and subsequently advanced and deposited it. In western Germany the Poznań moraine normally forms no marked ridge but consists of enormous quantities of sand and gravel (Woldstedt, 1925). The outwash, of course, indicates that much ice melting took place and that the halt after the readvance was due to a large supply during rapid wastage. North of Berlin the morainic line is poorly developed. East of

the Obra River, in central Poznań, the moraines are again marked, frequently forming long, steep ridges that locally rise a few hundred feet above their surroundings (Korn, 1914, pp. 480, 481). They primarily consist of bouldery and gravelly sand, though gravel also plays some part. Extensive wash plains are not common.

The northernmost morainic line, the Pommeranian, or the Great Baltic moraine, in western Poland and eastern Germany runs a considerable distance north of the Poznań; but in western Germany it is only a little inside, and in Schleswig-Holstein it merges with it. In Poland and eastern Germany a number of stadial moraines occur in the intervening region. The ice front apparently first withdrew a considerable distance and then advanced, especially in western Germany. Some of the moraines south of the Great Baltic morainic system are well developed and apparently represent marked halts. The time between the formation of the Poznań and the Pommeranian morainic lines, therefore, must have been long.

The Pommeranian moraine is topographically by far the most marked morainic line in northern Germany. In the western Baltic region it is especially marked and consists predominantly of till, while sand and gravel play little or no part (Woldstedt, 1925). In Pommerania it is distinguished by stony drift accumulated in hummocks and ridges with numbers of glacial lakes and ponds and with sand plains on the distal side.

The uncovering of the German Baltic region seems to have occurred with only few and brief interruptions, as moraines are scarce.

### THE DANISH ISLANDS AND SOUTHERN SCANIA

From the moraine in eastern Jylland, the correlative of the Pommerania moraine in Germany, the ice border withdrew eastward (Madsen, 1919, p. 111; Milthers, 1918, 1922; cf. hypothetical paper by Steensby, 1925, in which maps by Milthers and Madsen are reproduced; Storgaard, etc., 1925; Antevs, 1925b, p. 90). Moraine fragments on the Jylland east coast,

in Samsø, and in northern Zealand mark halts and small oscillations. If the retreating ice stream is identical with the older northeast ice in Scania, the Danish Islands may have become nearly or entirely ice-free and part of western Scania may also have become uncovered. Subsequently the ice readvanced from the southeast. Ice lobes pushed northward in the Danish Sounds and the Öresund and expanded over large parts of the islands to the northernmost of the younger morainic lines indicated on the map. In Scania the ice may have advanced to the region northeast of Romeleåsen (Munthe, 1920, pp. 65, 108). Then retreat followed. According to Munthe finally the southern and central parts at least of Scania were ice-free, and the temperature was comparable to that in the northern part of central Sweden at the present time. It follows that the Danish Islands would have become entirely uncovered. However, it seems probable to the present writer that the fossiliferous deposit at Robertsdal in southern Scania, on which this view is based, is wrongly dated, being, in fact, younger, an Alleröd bed. If this be the case perhaps only a comparatively short ice retreat took place before the new advance set in. During this latter advance a Baltic ice lobe extended to northern Öresund and spread over the coastal regions of eastern and southern Zealand and of western and southern Scania. At the same time the ice advanced from the northeast to a little beyond the northwest-southeast diagonal of Scania.

From the early part of the subsequent retreat there are some morainic lines in southern Scania. Soon the melting may have become rapid, for in all probability this is the time of the formation of the Alleröd beds of Denmark and southwestern Scania, whose flora and fauna indicate a temperate continental climate. During the Alleröd period the ice border retired beyond Bornholm and probably to or somewhat northeast of the northwest-southeast diagonal of Scania. Subsequently, as the temperature fell, as recorded by arctic plants in the massive clays on top of the Alleröd beds, the ice readvanced to Robertsdal, seven miles north of Ystad, provided the warm-climate bed there is an

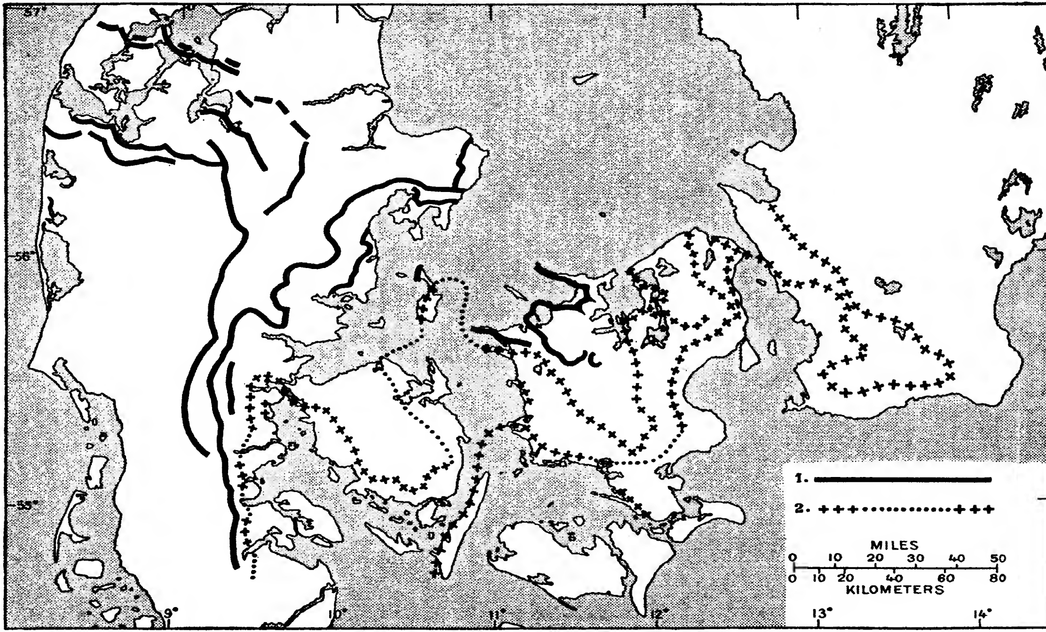



FIG. 28—Halts in the retreat of the last ice sheet in Denmark and Scania. Full-drawn lines (1) mark halts during the first retreat; crossed and dotted lines (2) mark limits of advances of the ice edge—crosses indicate definite, dots indefinite positions. (After N. V. Ussing, Axel Jessen, Vilhelm Milthers, Poul Harder, Victor Madsen, and Gerard De Geer; from Madsen, 1919, p. 111; northern Zealand from Milthers, 1927, p. 169).

Alleröd deposit, but it did not reach Bornholm. The later recession at first may have been slow. A local clay deposit with about 1500 varves at Rörum, six miles north-northwest of Simrishamn, perhaps means that the ice edge remained practically stationary for a corresponding number of years.

The oscillations of the ice border and the moraines show that the waning of the ice from the Danish Islands and southwestern Scania occupied a long time. Thick local deposits of varved clays in the flat land indicate the same thing, for although representing many hundreds of years several of these deposits must have been formed in small lakes ponded on all sides by ice, whose edge thus must have remained stationary for hundreds of years. In all, the uncovering of the belt between the Great Baltic, or the Pommerania, moraine and northeastern Scania must have taken many thousand years, probably some 10,000 to 15,000.

The formation of the moraines of the younger northeast ice and the low-Baltic ice in Scania (which moraines, however, are not to be correlated with the Great Baltic moraine in Germany) is dated by De Geer (1926a, Pl. 3) at about 9500 years before the bisection of the ice remnant at Stugun (63° 10′ N., 15° 35′ E.) in northern Sweden, the event which, according to the Swedish geochronology, inaugurated the postglacial age.

### SWEDEN AND FINLAND

From northeastern Scania and northward the withdrawal of the ice front is determined by means of the varved clay, but no datings of successive positions of the ice border are given for southern Sweden in De Geer's (1925, Pl. 2; 1926, Pl. 3; cf. 1914, Pl. 1, reproduced with Antevs, 1925b, p. 88) latest papers. Aside from a few brief retardations and halts marked by morainic belts (see Munthe, 1910) the mean annual rate of the retreat increased from 89 m. (292 feet) in northeastern Scania to about 150 m. (500 feet) south of Lake Vättern. The rate from Scania to the Fenno-Scandian moraines on Lake Vättern averaged some 120 m. (400 feet) a year, and the recession this distance took about 2000 years.

The great moraines crossing central Sweden and southernmost Norway and Finland—the finiglacial moraines, the Ras, and the Salpausselkäs, as their different parts are called—on the whole represent two marked halts separated by slow retreat, though they locally are divided into a number of small ridges. In Finland the first halt represents 225, the retreat 251, and the second halt 183 years (Sauramo, 1918, pp. 23, 35).[1] After rapid recession during 100 years retardation occurred for another 100 years. If this latter is included in the morainic belt this will represent 860 years (Sauramo, 1923, Pl. 8; 1926, p. 65). Later the decay of the ice quickly became very rapid. During the first 500 years after the withdrawal from the Third Salpausselkä the rate in western Finland averaged 380 m. (1247 feet) a year, the uncovered belt being 190 km. (118 miles) wide. As the region was then submerged the high rate was partly due to calving. In the corresponding zone in eastern Sweden the mean annual rate amounted to 400 m. (1312 feet), as the ice border withdrew from its position at the drainage of the Baltic Ice Lake at Mt. Billingen, now selected as the inauguration of the finiglacial epoch, up to Stugun (63° 10′ N.), a distance of 430 km. (267 miles) in 1073 years (De Geer, 1925, pp. 8, 17, Pl. 2; 1926a, Pl. 3). However, this latter rate is perhaps too large, for even if the drainage of the Baltic Ice Lake took place 292 years after the ice left the Second Salpausselkä, De Geer's figure for the finiglacial epoch is about 300 years smaller than that obtained by Sauramo (1926, p. 65).

### THE ALPS

The maximum extent of the last, the Würmian, ice cap and two important halts in its earliest retreat are recorded in the northern foreland of the Alps by three morainic lines (see Troll, 1925, p. 252). After the formation of this peripheral morainic belt, the ice border retreated without marked interruption until at length a halt occurred. It is called the Ammersee stage,

[1] The duration of the second halt is in Ramsay's (1924, p. 28) opinion unknown.

or Würm $\alpha$. After a renewed and long uninterrupted retreat to within the Alps another halt took place called the Bühl stage, or Würm $\beta$. Two additional halts of still later date in the ice recession are the Gschnitz, or $\gamma$, stage and the Daun, or $\delta$, stage, both of which are recorded by moraines throughout the Alps (Troll, 1925; A. Penck, 1925a, p. 359).

### Correlation between the Ice Retreat in Northern Europe and in the Alps

Although the waning of the last north-European ice sheet must have taken place contemporaneously with and corresponded to that of the Würmian glaciers in the Alps, the correlation of the stages of retreat in the two areas is not fully clear. The outermost Würmian moraines may be correlated with the Brandenburg moraines in northern Germany (Gagel, 1914; Woldstedt, 1925a, p. 184; A. Penck, 1925a, pp. 360, 366). The whole peripheral morainic belt in the Alps, then, probably corresponds to the Brandenburg and Poznań stages, and the Ammersee moraine to the Pommerania, or Great Baltic, moraine. The Bühl moraine may be correlated with one of the morainic lines in the Danish Islands, perhaps most likely with the youngest Danish line and its continuation in Scania. The Gschnitz stage perhaps corresponds to the Fenno-Scandian moraines, as held by Penck, for there is no marked halt in southern Sweden north of Scania that can come into consideration, and the most recent halt in Denmark and Scania is an unlikely correlative, as the distance between the Bühl and the Gschnitz is long. If so, the Daun stage, though marked throughout the Alps, lacks known correlative in Scandinavia.

There is also another consideration that makes the correlation of the three latest Alpine stages somewhat doubtful. Where the cooling effect of the ice did not make itself too strongly felt, the summer temperature may have risen so as to equal that of postglacial time long before the ice had been reduced to its present quantity. The melting of the ice took a long time, and it is practically sure that the huge ice sheets did not disappear

till a long time after the small ice caps or glaciers had either disappeared or had become reduced to their present size. Therefore, the Gschnitz stage which represents about one-half as much lowering of the snow line as do the outermost moraines of the Würmian glaciation, perhaps does not correspond to the Fenno-Scandian moraines, which lie about halfway between the limit of and the center of the Scandinavian ice sheet, especially as the ice when its border had retired to the Fenno-Scandian moraines was much thinned out and its mass only a fraction of what it had been during the greatest extent.

## Probable Correlation between the Ice Recession in North America and in Europe

Independent time estimates in North America and in Europe show that the last ice sheets in the two continents disappeared at the same time. The change from growth to shrinking of the ice sheets, caused by the most pronounced alteration of the climatic conditions that has taken place in the late-Quaternary age need not be supposed, however, because of local conditions to have taken place quite simultaneously at all points.

The ice melting was largely determined by the summer temperature. However, since the shrinking of the ice sheets was due to excess of wastage over supply, the same summer temperature may have had very different results according to topography, quantity of snowfall, etc. Large valleys facilitating ice flowage may have caused expansion of the ice, while in the same climatic area the ice melting normally was greater than the supply. Furthermore, different conditions of the sedimentation of the clay not seldom so influenced the thickness of the varves as to make them worthless for correlation.

In regard to widely separated regions, like North America and Europe, still other complications must be taken into account, viz. the uncertainty as to whether the same climatic changes took place and whether the yearly relative summer temperatures agreed. The climatic conditions in late-glacial time were so different from those in postglacial age and at present that

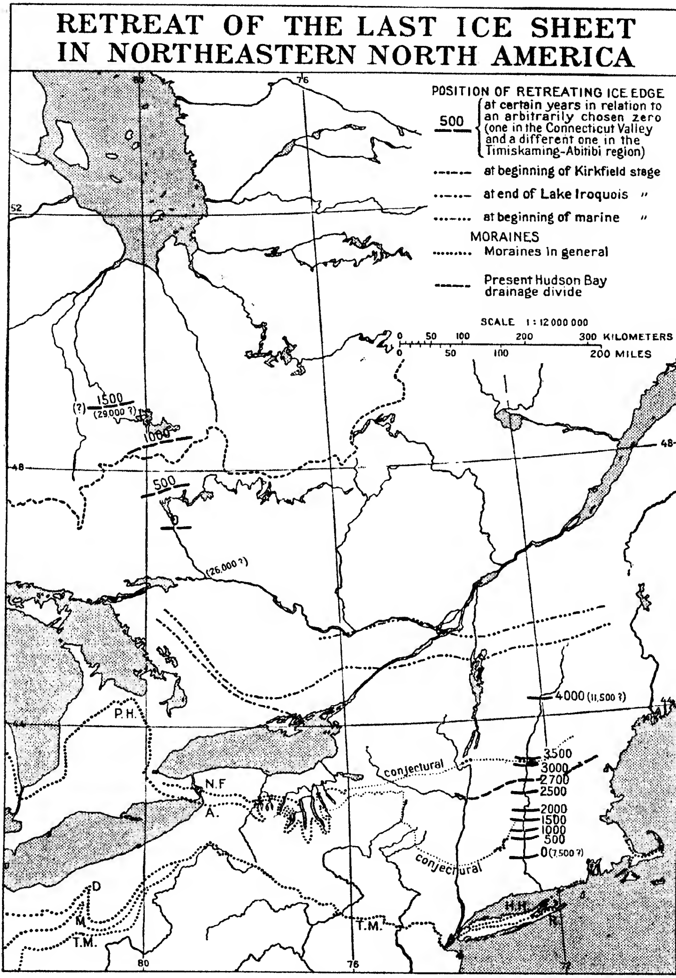



FIG. 29—Retreat of the last ice sheet in northeastern North America. (Moraines in the eastern states from various sources; moraines in the Middle West from Leverett and Taylor, 1915, p. 62; and moraines between the lakes from Taylor, 1924a.) R.—Ronkonkoma. H. H.—Harbor Hill. T. M.—terminal moraine. M.—Mississinawa. D.—Defiance. A.—Alden. N. F.—Niagara Falls. P. H.—Port Huron.

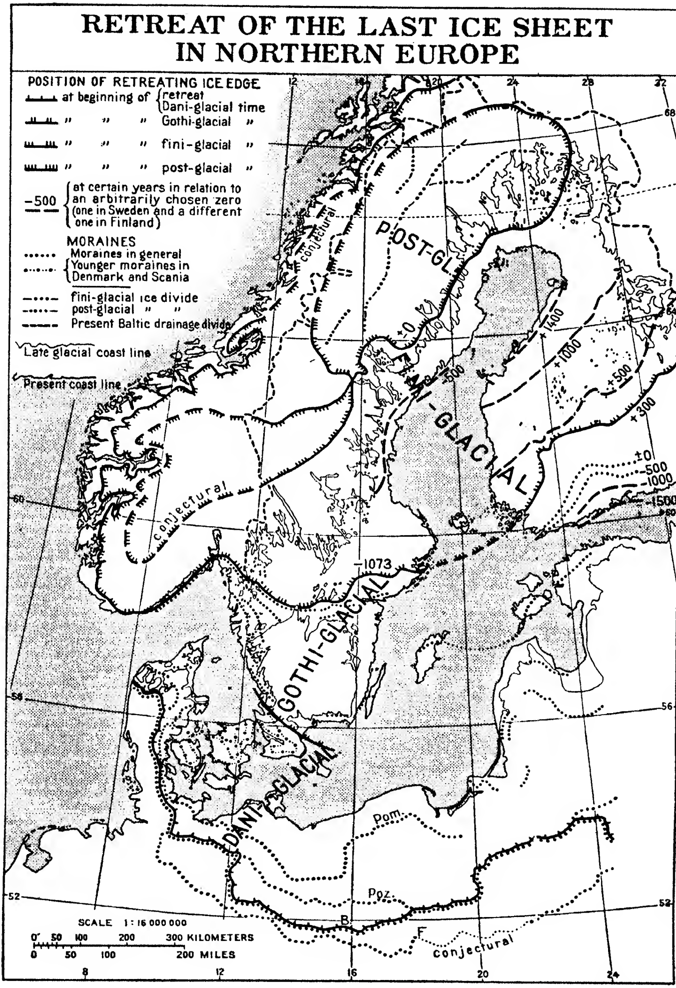



FIG. 30—Retreat of the last ice sheet in northern Europe. (Base map and stages in Sweden and Norway after De Geer, 1925. Stages in Finland and correlation with Sweden after Sauramo, 1923 and 1926. Moraines in Denmark after Madsen, 1919, and Milthers, 1922. Moraines in Germany after Woldstedt, 1925a. F., B., Poz., and Pom. designate the Flaming, Brandenburg, Poznań and Pommerania moraines.)

application of the knowledge of the latter to the time of the ice retreat offers difficulties. However, the postglacial temperature maximum and the climatic stress during the fourteenth century are recorded in both continents. There is at present a marked relationship between the climate of eastern North America and western Europe (Hildebrandsson, 1914, p. 4; Behler, 1922, p. 29). The whole North Atlantic area normally at one time undergoes a strengthening or a weakening of the atmospheric circulation coinciding with temperature departures in eastern North America and western Europe. It is thus highly probable that in late-glacial time the slower and more marked climatic changes in the two major areas of glaciation corresponded, while the yearly relative summer temperature, the ice melting, and the annual clay deposition cannot be assumed to have agreed.

Agreement between clay curves in itself does not necessarily prove contemporaneity, for the variation of the curves is rather limited and the correspondence may be due to recurrence of periods. Only such varve connections are valid as are based on curves showing good and persistent agreement, based on measurements from regions that were climatically similar in late-glacial time and that for good reasons can be assumed to have become ice-free at the same time. (See also Antevs, 1925b, p. 83; Milthers, 1927; cf. De Geer, 1926a.)

The exceedingly slow uncovering of southern Ontario was followed by a moderate rate of recession inaugurated by the uncovering of the Mattawa Valley 10,000 to 10,500 years ago according to the Niagara Falls chronology. The prolonged oscillations and the halts of the ice border in the Danish Islands and southern Scania ended by final retreat 13,500 to 14,000 years ago according to the Swedish clay chronology. The epochs of slow ice retreat or no retreat at all, so far as can be judged, were about equally long in both regions. The belts of retardation are of similar width and lie in similar situations within the glaciated areas. Each zone is the only one of its prominence in either area. Accordingly, the belts almost surely correspond to each other. As a consequence the Timiskaming region and

southern Sweden, in both of which the recession was rather rapid, also correspond. On the other hand, it is not clear whether the readvance at St. Johnsbury, Vt., or some other oscillation corresponds to the Great Baltic, or Pommerania, stage.

The correlation of earlier marked halts in the withdrawal of the ice front also meets difficulties. It is not certain that the moraines on Long Island and in Brandenburg correspond, but it is probable. The correlative of the Poznań moraine may be either the halts at Middletown, Conn., or one of the earlier halts in southeastern Massachusetts, either the Defiance or the Mississinawa morainic systems, most probably the Middletown-Defiance series. If this be so, the ice oscillations at Amherst and at Lake Winnepesaukee and their correlatives in the Great Lakes region correspond to moraines in northern Germany that were partly or wholly overridden by the advance forming the Great Baltic moraine. The longer oscillation in Europe may have been largely due to the fact that the Baltic Basin favored flowage of the ice.

The uncovering of the Timiskaming region, as has been said, corresponded pretty surely to that of southern Sweden. The annual mean rate of retreat in the former area was 515 feet (157 m.) and in the latter 400 feet (120 m.). As to situation, the belt of oscillation of the ice border at Iroquois Falls-Cochrane bears the same relation to southern Ontario that the Fenno-Scandian moraines bear to southern Scania and Danish Islands. The Iroquois Falls-Cochrane and the Fenno-Scandian stages are both the first marked halts north of the most important belts of retardation. However, while the whole Fenno-Scandian morainic belt represents 860 years, the first of the three observed oscillations in northern Ontario alone represents more than 670 years and probably some 1300 years, and the whole belt of oscillation represents perhaps more than 2000 years. Thus, while the first-mentioned conditions suggest that the belts are correlatives the lack of correspondence in length of time points in a contrary direction. It seems possible that the early part of the first oscillation in Canada coincided in time with the Fenno-

Scandian halts and that the subsequent temperature rise in northern Europe did not affect North America. It is accordingly held that the rapid decay of the European ice after its border had left the Fenno-Scandian moraines was due to the breaking in of the Gulf Stream into the Norwegian Sea (Enquist, 1918, p. 107). If this be so the border of the Labrador ice may have stood somewhat north of Cochrane when the European ice had shrunk to mere remnants.

Thus there was correspondence between the ice retreat in North America and in Europe in several of the larger features, but topographic and climatic differences seem to have limited the agreement. Since the correspondence was not perfect even in the larger features, agreement in the smaller features and in details, such as relative summer temperatures and varve graphs, cannot be expected.

Our present knowledge of the geologic history of the two areas does not permit any other correlation. If the one outlined is correct and the estimates of the time represented by zones in which the ice retreat is not chronologically determined are fair, the last ice sheets had their greatest extent and began to wane about 40,000 years ago. This figure may be less than 10,000 years too large or too small—a fact of importance because of the interest that has recently sprung up in the absolute Quaternary chronology (Köppen and Wegener, 1924; Soergel, 1925a).

# CHAPTER IX

## DISTRIBUTION OF VARVED CLAY IN THE AREAS HERE TREATED

This chapter and the next two present and discuss material that has not been covered by earlier reports on ice recession. In some instances the observations were made as far back as 1921; but in most cases they are more recent. The several areas visited are so different as regards terrain and late glacial history that a brief description of each one seems desirable, to make clear why the varved clay records found in the several fields are so variable in value, and whether further study promises results or not.

In the peripheral belt of the glaciated area from the Delaware River eastward to Cape Cod varved glacial clay has scanty distribution. The small quantity of clay here as well as in certain other regions is a remarkable phenomenon. The chief reason may be that the expanded parts of the valleys were first filled by protruding ice tongues and that later the glacier mud, which did not settle quickly, to a great extent passed through small water bodies even though stagnant and was carried to the sea or to large and deep lakes. Furthermore, the clay that actually occurs in the area here considered is accessible only in a limited number of valleys. Practically every valley that seemed at all promising has been visited. (Fig. 5, p. 108.)

In southern New York, between the Delaware and the Hudson, some brickyards on late-Quaternary clay were formerly operated (Ries, 1900, pp. 731–733; map). At present the clay stands under water or is for other reasons inaccessible. In northern New Jersey, where varved clay also is to be found in the lowest, swampy parts of many valleys (Ries, Kümmel, Knapp, 1904; Salisbury, 1902), it is exposed at but few places. Along the Hudson late-glacial clay seems to be limited to districts in which measurements have been made, since the valley has been care-

fully explored. On Long Island varved clay is scarce. At Greenport, on the northeastern peninsula, boulder clay, which shows no indication of being crumpled varved clay, is used for brickmaking. The clay at the brickyard two miles north of Farmingdale may have been a varved clay overridden and thoroughly kneaded by the ice. It is overlain by gravel, at least two feet thick.

The river valleys in western Connecticut seem to have been too narrow to allow the deposition of silt and clay. In the middle of the state varved clay is rather widely distributed (Loughlin, 1905; F. Ward, 1920; U. S. Geol. Survey Water Supply Papers 374, 397, 449, 466, 470, 537). The region has been carefully investigated. In Rhode Island and southeastern Massachusetts late-glacial clays occur at various places (Shaler, Woodworth, Marbut, 1896), the chief areas being those from which sections have been obtained.

In northernmost New England and southernmost Canada, i.e. in the area from the upper Connecticut River and the St. Francis River westward to Lake Ontario and Georgian Bay and from Lake Ontario northward to Maniwaki, 70 miles north of Ottawa, little varved clay occurs except in the valleys where measurements have been made, for practically all promising regions have been carefully investigated and all observed exposures measured (see also Antevs, 1925b, p. 57; Jacobs, 1926). The extreme scarcity of varved clay on Lake Champlain and in southern Ontario is remarkable. The clay must have been washed down into the deepest parts of the valleys and the lakes during the oscillations of the land. (Fig. 2, p. 95; Fig. 4, p. 99.)

In northern Ontario and Quebec varved clay has very extensive distribution up to a line running about east and west through Iroquois Falls, i.e. to the limit of the first readvance of the ice in the region (see also Antevs, 1925b, p. 58). It here covers practically all the low land. As the country except for narrow strips along the railways is a wilderness, only natural exposures are to be found. These are furthermore very infrequent because of the flatness of the land, because of the generally low or well-graded river banks covered by a dense growth of trees and bushes down

to the water's edge, and because of the recent damming up of several rivers to develop power. (Fig. 23, p. 142; Fig. 22, p. 147.)

North of Iroquois Falls the varved clay is covered by boulder clay, usually 20 to 25 feet thick, and is therefore seldom accessible. Only the Frederick House River between Frederick House Lake and locality 135, eight miles north of the lake, offers good exposures in the deep trench formed after the river in 1909 was diverted by a reckless prospector (Knight, etc., 1919, p. 42). The water previously flowed over a shoulder of bed rock and formed a fall 46 feet high. Just to the northeast of the fall there was a narrow barrier of deep soft boulder clay. The water was diverted into a small cutting in this clay embankment and soon carved a gorge down to the river level below the dead fall. (Fig. 26, p. 146.)[1]

The rivers traversed by canoe are the following: Villemontel River from the road straight south of Villemontel station to the southern part of Kewagama Lake; La Sarre River, Lake Abitibi, Abitibi River, and Kanasuta River from locality 118, five miles north of La Sarre, to the southern part of Dasserat Lake; Black River from Matheson to High Falls near locality 119; Abitibi River from Iroquois Falls to a point three miles north of the Transcontinental Railway; Frederick House River from the central part of Night Hawk Lake to a point a few miles north of the Transcontinental Railway; Buskegau River from the railway upstream to five miles above the junction of the main branches; Driftwood River from the railway 25 miles upstream; Mattagami River from the rapids eight miles southwest of Timmins to the Transcontinental Railway; and Missinaibi River from Mattice station 15 miles upstream.

[1] It is of interest to compare this drainage with that in 1796 of Lake Ragunda (63° 10′ N., 16° 20′ E.) in northern Sweden (Ahlmann, 1924, pp. 50, 112). Lake Ragunda was an expansion of the Indalsälven caused by damming of the preglacial river channel by glacial clay. The water flowed across a rock sill and formed a large fall. As the fall prevented log driving and boat communication between the regions above and below, it was decided to lead the river into its old channel. To this end a brook was diverted out on the clay barrier; and finally the lake was catastrophically drained in the course of about two hours, causing great damage along the river banks below. In this case chronological material of still greater importance was exposed, as the lake beds were deeply trenched.

# CHAPTER X

## DESCRIPTION OF THE SECTIONS AT THE LOCALITIES STUDIED

In this chapter are detailed descriptions of sections studied and measured at 160 localities. These supplement earlier reports of studies in New England, the Hudson Valley, and Canada (Antevs, 1922, 1925b). Of these new localities, 30 lie in New Jersey and the Hudson Valley, 26 in southern New England, 33 in northern Vermont, and 71 in Canada. They bring the total number of localities studied to 150 in New England, 30 in New Jersey and the Hudson Valley, and 174 in Canada. Many other localities were visited and studied during the progress of the field work but afforded little or no significant information.

Nearly all locations are shown on the accompanying maps. Number 92 is assigned to the first of the new localities because 91 had previously been described from the New England region (Antevs, 1922). Similarly, but with independent numbering, in Canada, the first locality here is Number 104, a locality previously studied and designated by that number, but remeasured in 1925; and numbers 105 to 174 are assigned to the new localities.

In each case the beds in a section are enumerated from the top downward. The varves are numbered as in previous publications. In the case of the central New England series, and measurements connected with it, the numbering is based on the arbitrary number 3001 as the lowest varve reached in the section at Hartford, Conn. (Antevs 1922, p. 49). The series identified at New Haven and Haverstraw, but not connected with the long New England line, is numbered independently. Local numbers are used, likewise, for unconnected series at localities in Long Island, Rhode Island, southeastern Massachusetts, etc. The same system is used in Canada, where a limited amount of group-

ing is possible from connections. The longest series here is that of Timiskaming, with more than 2000 varves.

Parentheses indicate that the varves enclosed by them have been used only in correcting, not in constructing the normal curve—as, for example, under locality 93 "(886–908)." The date when each measurement was made is given after each location.

## Localities in New Jersey

These are shown on Figures 6 and 7, p. 110. Essentially the same localities as our nos. 92–99, measured in 1921 (Antevs, 1922, pp. 7, 8), 1922, and 1924, were in 1923 and 1924 studied by Reeds (1926). Plate I.

92.—3¼ miles S of Hackensack, 1 mile SSW of Little Ferry, the southernmost brickyard on the west side of the Hackensack River, 200 yards S of the office building. 1924.

6 feet sand.
7⅔ feet silty clay. Varves not distinguishable.
5 inches silty clay, about 5 varves.
6½ inches clay. Varves too thin to be determined.
2 feet 2 inches clay, about 200 more or less indistinct varves.
3¼ feet stiff clay, varves 1275–1490.
8 inches stiff clay. Varves very thin; about 65.
Probably about 70 feet to substratum, since the clay at the brickyard is found by borings to be 85 feet deep (Salisbury, 1902, p. 510).

Series measured: 1275–1490.

93.—Same brickyard as No. 92, 50 yards W of the office building. 1924.

Many feet of sand and clay removed.
2½ inches silty, light-red clay, sharply set off against the underlying clay, varves 947–952.
6¾ feet stiff, gray to red-brown clay, varves 645–946. Varves 873–878 have bright-red winter layers.
Far to bottom.

Series measured: 645–952, 659–885, 909–951, (886–908).

94.—Same brickyard as Nos. 92 and 93, 70 yards W of the office building. 1924.

Many feet of sand clay removed.
7⅓ feet stiff, gray to red-brown clay, varves 593–835.
Far to substratum.

Series measured: 593–835, 593–775, 639–675.

95.—Same brickyard as 92 etc., 250 yards W of the office building. 1922, 1924.

5 feet sand.
3 feet sandy clay.
4 feet clay, varves indistinct.
5 feet rather stiff red-brown clay, varves 1101–1392. Clay at the base silty and light red, sharply set off against the underlying clay.
1⅔ feet clay, at the base somewhat silty and light-red, but soon growing stiff and red-brown to gray upward. A great number of very thin clay layers, so that the varve limits are not distinguishable. Probably between 75 and 200 varves.
2½ inches somewhat silty, light-red clay, varves 947–952. Very distinctly set off against the underlying clay.
1 foot stiff, gray-brown clay, varves 886, 946.
Far to bottom.

Series measured: 886–952, 1101–1290, 1101–1162, 1101–1128, 1164–1392.

96.—Brickyard ½ mile SW of Little Ferry. 1921, 1922.

5 feet sand.
8 feet sandy, almost massive clay.
2 feet silty, at the base sandy, clay, 55 varves. The lowest varves distinctly set off from the clay beneath, and record drainage.
5 feet rather stiff, indistinctly varved clay, about 400 varves.
3¾ feet stiff, gray to red-brown clay, varves 1241–1490.
According to information about 70 feet to substratum.

Series measured: 1241–1490.

97.—Brickyard ⅓ mile W of Little Ferry. 1921, 1922, 1924.

2 feet sand.
3 feet leached clay.
6 feet silty clay, varves difficult to distinguish.
2 inches silty, light-red clay, distinctly set off against the underlying clay, varves 947–952.
3½ feet fat, gray-brown clay, varves 700–946. Varves 873–878 have brightly red winter layers.
4 inches fat clay, varves difficult to distinguish, varves 673–699.
15½ feet mostly stiff, above 570 silty, gray-brown to red-brown distinctly varved clay, varves 114–672. Varves 185 and 283 begin series of varves with winter layers of bright-red color, which gradually fades.
½ foot disturbed clay.
Till.

Series measured: 114–300, 273–327, 275–471. 319–658, 340–658, 607–672, 700–952.

98.—Brickyard ½ mile NW of Little Ferry. 1922.

8 feet sand.
1¾ feet sandy clay, varves 593–636. Surfaces of the varves undulating.
2⅓ feet somewhat sandy clay, varves 523–592.
8 feet distinctly varved clay, varves 244–522.
According to information 6 to 8 feet to substratum.

Series measured: 244–592, 351–414, 523–554, (592–636).

99.—Brickyard 1 mile NNW of Little Ferry and 1½ mile S of Hackensack. 1921, 1922.

7 feet sand.
7 feet crumpled clay.
8 feet fat clay, varves 1–321. Varves 75–93 and 185–321 are red; the others gray or gray-brown. Varve 1 is bottom varve.
Till.

Series measured: 1–56, 1–37, 9–64, 9–73, 21–93, 53–73, 53–73, 76–321, 123–271, 165–218.

100A.—5 miles N of Hackensack, Oradell, bluff on the W side of Hackensack River, 100 yards N of the station. 1924.

3½ feet sand.
2½ feet silty clay.
6 feet silty clay, about 140 varves. In some varves sand, so that thicknesses are frequently not characteristic. Color gray-brown to red.
River level. Depth to substratum probably great (see Salisbury, 1902, p. 616).

Series measured: Series not connected with any other. 140 varves; no curve to be published.

100B.—150 yards NE of locality 100A, the east river bank, 25 yards below the dam. 1924.

5 feet sand.
2½ feet sandy clay, 60 varves. Part of the varves consist of yellow-brown sand and strongly red clay laminæ—winter layers. Thicknesses poor.
River level. Depth to bottom unknown.

Series measured: Series not connected with any other. 60 varves; no curve to be published.

101.—Brickyard 1½ miles NW of Little Falls. 1922.

2½ feet clayey sand.
2½ feet sand.
2½ feet thoroughly contorted clay with scattered boulders, almost certainly till; overriding of the clay by the ice. Salisbury (1902, p. 526; photograph in Ries, etc., 1904, Pl. 16, p. 128) found here typical till.
7½ feet somewhat silty gray-brown, distinctly varved clay, 222 varves. Measurements made on clay blocks with almost undisturbed lamination occurring in the otherwise contorted clay deposit.
Probably insignificant depth to bottom; subjacent gravel at this level close by.

Series measured: This series is connected only with that at locality 102. Little Falls 1–222, 78–149, 89–130.

102.—Brickyard ⅓ mile N of Mountain View. 1922.

7 to 10 feet sand.
7 feet varved clay, contorted in some zones so as to be almost massive.
2 feet almost homogeneous clay with scattered stones, likely till.
1 foot clay, varves 90–111.
2 inches slidden clay.
8 inches clay, 17 varves.
Contorted clay. According to information 4 to 5 feet to bottom.

The exposed clay deposit may represent about 400 varves. The clay was overridden by the ice at least once, more probably several times.

Series measured: Series connected with that at locality 101. (90–111).

### Localities in the Hudson Valley

These are shown on Figure 8, p. 111 and Figure 9, p. 113. Plates I–III.

103.—Haverstraw, N. Y., brickyard ⅔ mile NE of the station, near the Hudson River at the northern edge of the town. 1921.

Surface about 5 feet above the Hudson and sea level.
5 feet sand.
7 feet silty clay.
23 feet silty, gray clay, beautifully varved, varves 483–634.
Far to substratum.

Series measured: The series at Haverstraw is connected with that at New Haven, Conn. New Haven 483–634.

104.—Clay pit 300 yards NW of locality 103. 1921.

Sand and gravel.
2 feet crumpled clay.
12 feet gray, silty clay, varves 395–489.
Far to substratum.

Series measured: New Haven 395–489.

105.—Clay pit 300 yards NW of locality 104. 1921, 1922.

5 feet disturbed silt and sand.
4½ feet sandy silt, varves 708–732. Surfaces of some varves wavy.
13 feet clayey and sandy silt, varves 641–707.
44 feet silty clay, varves 331–640. Lamination beautiful, with gray summer layers and dark, shining winter layers.
Bottom of clay pit. Far to substratum.

Series measured: New Haven 331–732, 397–422, 579–619, 588–722, 698–732.

106.—Clay pit 200 yards N of locality 105. 1921.

3 feet sandy silt.
2 feet coarse silt, 13 varves. Ripple marks; thicknesses not good.
1½ feet contorted silt.
2½ feet coarse silt, 7 varves. Ripple marks.
14½ feet silty clay or clayey silt, varves 602–697.
Several feet not measured, being inaccessible.
4¼ feet silty clay, varves 488–520.
Far to bottom.

Series measured: New Haven 488–520, 602–697.

107.—1¼ mile N of Haverstraw station, clay pit 300 yards NE of West Haverstraw station. 1921.

8 feet gravel, locally sand, with boulders large as a head. Locally in the large pit 3 feet silt with gravel and boulders —till (?).
9½ feet, disturbed largely, massive silt.
1¾ feet sandy silt, varves 548–577.
A few inches disturbed silt.
15 feet sandy silt, varves 143–513. Summer layers consist of silt and fine sand. The greasy winter layers occasionally represent one-half of the varve thickness. Thicknesses not quite good.
Bottom of clay pit. Depth to substratum unknown.

Series measured: New Haven 143–513, 548–577.

108.—Clay pit 300 yards N of locality 107. 1921, 1922.

8 feet sand and gravel.
8 feet disturbed silt and clay.
3 feet silty clay, varves 519–569.
Contorted zone.
11⅔ feet sandy and silty clay, varves 265–494.
1 foot sandy silt, disturbed at top. A single drainage varve.
19½ feet sandy and silty clay, varves 21–263. Summer layers consist of sandy silt. The greasy winter layers sometimes constitute as much as one-half of the thicknesses of the varves. Thicknesses fairly good.
3 inches crumpled clay.
1⅓ feet sandy and silty clay, varves 1–15.
Disturbed sandy clay at bottom of clay pit. Depth to substratum unknown.

Series measured: New Haven 1–15, 20–97, 21–87, 57–156, 87–499, 150–392, 183–262, 266–301, 267–366, 395–544, 395–415, 417–543, 519–569.

109.—Haverstraw, 2 miles N of the station, brickyard on the Hudson at Grassy Point. 1922.

1½ feet gravel.
4 feet disturbed clay.
20 feet gray silt clay, varves 58–301.
3¾ feet varved sand, varves 38–57.
1½ inches crumpled sand.
2½ feet varved sand, varves 12–35.
Not far to bottom.

Series measured: New Haven 12–35, 38–301, 62–100, 162–193.

110.—Peekskill, 3½ miles S of the station, ¾ mile WNW of Crugers, brickyard on the Hudson at Georges Island. 1922.

The clay is quite disturbed. It usually reaches up to the level ground surface at or slightly above the high-water level of the Hudson. At places the varved clay is eroded, and the channels of erosion are filled with post-glacial clay, sand, and peat.

110.—*Continued*

1½ feet sand.
1½ feet sandy clay, postglacial.
1½ to 3 feet peat with tree trunks.
2 to 3 feet unvarved postglacial clay.
5½ silty clay; varve limits difficult to distinguish; varves 542–572.
17 feet silty, gray, distinctly varved clay, varves 443–541.
8 inches through sliding massive clay.
5⅔ feet silty, gray clay, varves 406–440.
Far to bottom.

The varve curves show excellent agreement with those at Haverstraw, but the varves are thicker.

Series measured: New Haven 406–440, 443–541, 542–572.

111.—Same brickyard as locality 110, 200 yards from 110. 1922.

Swamp flora with *Typha*.
1 inch peat.
1 foot pure massive clay, postglacial.
5 feet peat with tree trunks.
Some homogeneous, postglacial clay.
9 feet silty clay, varves 420–495.
4 feet crumpled and partly covered clay.
11½ feet silty, gray clay, varves 265–379.
Depth to substratum probably great.

Series measured: New Haven 265–379, (420–495).

112.—Dutchess Junction, clay pit 900 yards S of the station and 100 yards E of the railroad tracks. 1922.

At localities 112–114 and 116–120 the varve curves are good, but the thicknesses of the varves are mostly greater than at the other localities, so that only a part of the measurements have been used in constructing the normal curve.

10 feet sand.
3 feet varved silt.
6¾ feet gray, clayey silt or silty clay, varves 2923–2943.
5 feet disburbed clay silt.
5½ feet silty, gray clay, varves 2881–2900.
2 feet crumpled clay silt.
5½ feet clay silt, varves 2859–2875.

112.—*Continued*

8 inches slidden clay silt.
10⅓ feet clayey silt, varves 2828–2857.
Depth to bottom unknown.

Series measured: Hartford (Antevs, 1922, p. 49) (2828–2857, 2859–2875, 2881–2900, 2923–2943, 2912–2925, 2913–2922, 2827–2937).

113.—Dutchess Junction, clay pit 500 yards S of the station and 150 yards E of the tracks. 1922.

10 feet silt and sand.
6 feet sandy silt, varves 2918–2927. Varve surfaces wavy.
9⅓ feet silty, gray clay, varves 2877–2917.
8 inches contorted clay.
2½ feet silty clay, varves 2868–2875.
8 inches crumpled clay.
12¾ feet silty clay, varves 2828–2866.
5 inches slidden clay.
8 feet silty clay, varves 2806–2826.
Contorted clay at bottom of clay pit. Depth to substratum unknown.

Series measured: Hartford 2883–2911, (2806–2826, 2828–2866, 2868–2875, 2877–2882, 2912–2927).

114.—Dutchess Junction, clay pit 400 yards S of the station and 150 yards E of the tracks. 1922.

3 feet unvarved sand.
3 feet varved sand.
8 feet silty sand, varves 2929–2943.
4 inches disturbed sand.
7 feet sandy silt, varves 2903–2927.
6 feet sandy and clayey silt; not accessible.
1½ feet clayey silt, varves 2872–2881.
8 feet somewhat crumpled silt.
10 feet clay silt, varves 2832–2866.
8 inches contorted clay silt.
10⅔ feet gray clay silt, varves 2806–2835.
Crumpled clay silt at bottom of clay pit. Depth to substratum unknown.

Series measured: Hartford 2903–2927, (2806–2835, 2837–2866, 2872–2881, 2929–2943).

115.—Dutchess Junction, clay pit 150 yards E of the station. 1922, 1924.

7 feet sand.
8 feet varved sandy silt.
5¼ feet sandy and silty clay, varves 2900–2911. Varve surfaces wavy.
67 feet gray, distinctly varved clay silt, varves 2700–2899. Single varves disturbed locally.
Depth to bottom unknown.

Series measured: Hartford 2700–2810, 2704–2899, 2719–2773, 2851–2895, (2900–2911).

116.—Beacon, clay pit ¾ mile S of the station and 100 yards E of the railroad tracks. 1922, 1924.

5 feet clayey gravel.
1½ feet unvarved sand silt.
8 feet sand silt, about 15 varves. Layer surfaces wavy.
5½ feet quick silt, varves 2840–2864 (?). The 10 uppermost varves with undulating surfaces.
3 feet not accessible.
6¾ feet clay silt, varves 2780–2806.
3 feet disturbed clay silt.
5 feet disturbed clay silt, about 15 varves. A downslidden zone.
2 feet contorted clay silt.
6 feet silty clay varves 2767–2783.
2 feet disturbed silty clay.
13 feet silty clay, varves 2729–2759.
Bottom of clay pit. According to information about 40 feet to substratum.

Series measured: Hartford (2729–2759, 2737–2753, 2767–2787, 2767–2783, 2780–2806, 2840–2864 (?)).

117.—Beacon, abandoned clay pit 500 yards S of the station and 100 yards E of the tracks. 1922.

4 feet disturbed silty clay.
3½ feet silty clay, varves 2743–2752.
1 foot disturbed silty clay.
About 10 feet clay silt to till. Rock crops out close by.

Series measured: Hartford (2743–2752).

118.—Beacon, 2 miles N of the station, clay pit at Brockway, 200 yards E of the railroad tracks. 1921, 1922.

3 feet clayey gravel.
2 feet disturbed clay silt.
20 feet gray, silty, distinctly varved clay, varves 2843–2893.
2 feet crumpled clay silt.
5 feet silty clay, varves 2831–2840.
Far to bottom.

Series measured: Hartford (2831–2840, 2843–2893).

119.—Clay pit 100 yards N of locality 118. 1922.

Several feet contorted silty clay.
13⅔ feet silty, distinctly varved clay, varves 2812–2835.
Probably far to substratum.

Series measured: Hartford (2812–2835).

120.—Clay pit 400 yards ENE of locality 118. 1921, 1922.

2½ feet clayey gravel with cobbles.
1 foot leached, silty clay.
21 feet silty clay, varves 2806–2859. These overlap the underlying series in spite of the contorted zone, showing that there must have been horizontal sliding.
1 foot crumpled clay silt.
8 feet clay silt, varves 2799–2813.
Far to substratum.

Series measured: Hartford (2799–2813, 2806–2859, 2817–2855, 2828–2848).

121.—Newburg, 4¾ miles NNE of the station, clay pits 1 mile NNE of Roseton station. 1922.

15 feet gravel.
33½ feet gray, distinctly varved clay silt, varves 2990–3170.
1½ feet crumpled clay silt.
33 feet gray, distinctly varved clay silt, varves 2900–2988.
2 feet contorted clay silt.
5 feet clay silt, varves 2888–2897.
Fault.
12 feet clayey and sandy silt, varves 2865–2884.
Depth to bottom unknown.

121.—*Continued*

Series measured: Hartford 2900–2988, 2900–2937, 2906–2932, 2906–2932, 2990–3170, (2865–2884, 2866–2889, 2888–2897, 2889–2896).

## Locality on Long Island

122.—20 miles ENE of New York, 73° 38′ W., 40° 50′ N., Glen Head, clay pit ½ mile E of the station. 1924. Pl. IV.

4 feet clayey gravel.
6 feet sandy laminated but unvarved silt and silty sand.
7 feet clay silt. Varve limits very indistinct.
2¾ feet clay silt, varves 56–71.
1 foot clay. Varve limits very difficult to distinguish. Varves 31–55.
2¾ feet beautifully varved clay, varves 1–30.
Bottom of clay pit. Depth to substratum unknown.

Series measured: The series not connected with any other series. 1–71, 9–57, 14–40.

## Localities in Connecticut

These are shown on Figure 10, p. 115 and Figure 11, p. 116. Plates II–IV.

123.—New Haven, 3 miles NE of Union Station and 1 mile NE of East Rock, clay pit on the E side of the railroad to Hartford. 1921, 1923.

Ground surface slightly above high-tide level
3 feet peat.
5 feet fine gravel and sand.
5 feet sandy and silty clay.
17 feet red, beautifully varved clay, varves 316–544. Winter layers comprise one-third to one-half of the varve thicknesses. The clay is somewhat compressed, but the relative thicknesses of the varves are rather good.
Depth to substratum great (see F. Ward, 1920, p. 52).

Series measured: The series at New Haven is connected with that at Haverstraw, N. Y., and has the same numbering. New Haven 316–498, 403–528, 501–544.

124.—Clay pit 600 yards N of locality 123, on the E side of the tracks. 1921, 1922.

At this place the clay is much pressed and otherwise disturbed, so that only two out of several measurements proved to be of value.

3 feet peat.
2 feet gravelly sand with remains of plants.
3 feet sandy clay.
5 feet red, silty clay, varves 349–433.
2 ½ feet silty clay, not accessible.
4⅓ feet red clay, varves 233–308.
Depth to substratum probably fairly great.

Series measured: New Haven 349–433, (233–308).

125.—Clay pit 1¼ miles N of locality 124, on the E side of the tracks and 300 yards N of Quinnipiac flag station. 1921, 1922.

8 feet unvarved sand.
6 feet fine varved sand.
3 feet disturbed sandy clay.
13 feet red clay, distinctly varved, varves 324–502. Winter layers form one-third to one-half of the varve thicknesses
Disturbed zone.
2⅔ feet clay, varves 289–321.
Depth to substratum great.

Series measured: New Haven 324–502, 392–502, 398–502, 434–460, 467–502, (289–321).

126.—New Haven, clay pit 4 miles NE of Union Station and ⅓ mile E of the railroad to Middletown. 1922.

5 feet sand.
3 feet compressed clay.
8 feet red, silty clay, varves 241–387. Clay pressed so that varve thicknesses are not characteristic.
Depth to substratum fairly great.

Series measured: New Haven (241–387, 244–319).

127.—New Haven, 6 miles NE of Union Station, clay pit 1¼ miles S of North Haven. 1922.

6 feet sand.
4½ feet partly disturbed clay.
12 feet red, distinctly varved clay, varves 180–300.
Depth to bottom probably not great.

Series measured: New Haven 180–300, 202–227, 232–257, 241–274, 261–291, 261–273.

128.—Clay pit ¾ mile N of locality 127, and ½ mile SSE of North Haven. 1922.

4 feet sandy gravel.
½ foot sandy clay.
14½ feet red, distinctly varved, in the lower part somewhat sandy, clay, varves 189–244.
Depth to bottom probably not great.

Series measured: New Haven 275–344, 275–333, 291–333, 275–296, (189–286, 204–282, 220–252, 254–279).

129.—Middletown, 1¾ miles NNW of the station, clay pit 150 yards N of Newfield station. 1921.

7 feet leached clay.
9 feet distinctly varved, silty clay, varves 15–39.
Far to substratum.

Series measured: Newfield 15–39.

130.—Clay pit 150 yards N of locality 129. 1921.

3 feet sand.
5 feet partly disturbed varve clay.
5¾ feet distinctly varved clay, varves 14–37. The three top varves sandy; the others greasy.
According to information 15 feet to bottom.

Series measured: Berlin 14–37 (cf. locality 133).

131.—Clay pit 300 yards N of locality 129. 1922.

2 feet stony clay, probably till.
8 feet red, distinctly varved, somewhat pressed clay, varves 1–30.
Depth to bottom not great.

Series measured: Newfield 1–30, 5–29, 6–25.

132.—Middletown, 2½ miles NNW of the station, clay pit on the west side of the railroad track, Tuttle Brickyard No. 3. 1922.

2½ feet till—clay, gravel, cobbles.
3 feet contorted clay, 15 to 20 varves.
3 feet clay, partly pressed, 21 varves.
6 inches crumpled clay.
5½ feet beautifully varved clay, 32 varves.
Till, exposed to a depth of 5 feet.

133.—Berlin, 1¾ miles E of the station, clay pit just E of Beckley station and N of the railroad tracks. 1921.

3 feet fine sand and clay in alternating layers about ¼ inch thick—not varves.
3 feet massive clay with scattered stones—till (?).
4 inches clay, probably 3 varves.
4¼ feet silty clay, varves 11–32.
Close by—2⅔ feet silty clay, varves 1–9.
Probably far to bottom.

Series measured: Berlin 1–9, 11–32.

134.—Berlin, ¾ mile E of, clay pit just S of the tracks. 1921.

2 feet sandy and stony clay, probably till.
2 feet sand.
1 foot leached and disturbed clay.
5 feet distinctly varved clay, varves 1–30. Summer layers gray-brown; winter layers sharply red-brown.
Depth to bottom unknown.

Series measured: Berlin 1–30.

135.—Berlin, clay pit 500 yards ENE of the station, just N of the railroad tracks to Middletown. 1922.

3 feet till, i.e. clay with gravel and pebbles kneaded in.
2 feet distorted varved clay.
1½ feet silty clay, 1 varve.
8 inches disturbed clay.
4 feet distinctly varved clay, varves 10–33.
Inconsiderable depth to till that outcrops close by.

Series measured: Berlin 10–33.

136.—Berlin, clay pit 700 yards NE of the station, just E of the railroad tracks to Hartford. 1922.

2 feet sand and till.
1 foot pressed clay.
4 feet red, distinctly varved clay, 16 varves.
3 feet disturbed clay, about 10 varves.
Till.

Series measured: Berlin, 1–16.

137.—New Britain, 2¼ miles NE of the station, abandoned clay pit in the NW corner of the railroad running from New Britain to Hartford and the highway running eastward to Newington. 1921.

7 feet sand.
4 feet varved, sandy clay with lenses of sand.
2½ feet silty clay, varves 17–33.
2 inches disturbed clay, probably representing 1 or 2 varves.
5 feet red, silty clay, varves 1–14.
10 feet disturbed clay.
According to information about 10 feet to quicksand.

Series measured: New Britain 1–14, 17–33.

138.—Hartford, 7 miles S of the station, ¾ mile S of South Wethersfield station, bluff on brook on the west side of the highway. 1922.

12 feet till, i.e. crumpled clay with pebbles and boulders.
4¼ feet red clay, 12 varves.
Till, exposed 5 feet down to brook level.

Series measured: Not connected with any other. 1–12.

139.—Hartford, 5½ miles SE of the station, ½ mile ESE of Naubuc, bluff on Salmon Brook. 1922.

7 feet sand and gravel.
6 inches leached clay.
10 feet stiff, red-brown, distinctly varved clay, varves 2996–3121.
2 inches contorted clay.
1¾ feet clay, varves 2868–2891.
Contorted clay at brook level. Depth to bottom unknown.

139.—*Continued*

Series measured: Hartford (Antevs, 1922, p. 49) 2868–2891, 2868–2887, 2896–3121, 2897–3080, 2897–2911, 2940–2950, 2971–3017, 3007–3017.

140.—Hartford, 2¾ miles S of station, ¾ mile ENE of locality 1 (Antevs, 1922, p. 11), bluff on South Fork where it turns northward. 1922.

10 feet varved clay, not measured because of lack of time.
1¾ feet distinctly varved clay, varves 3162–3207. Curve shows excellent agreement with the published normal curve (Antevs, 1922, Pl. 1).
Disturbed zone. Two feet covered to brook level.

Series measured: (Hartford 3162–3207).

## Localities in Rhode Island

These are shown on Figure 12, p. 118. Plate IV.

141.—Providence, 6 miles S of the railroad station, bluff on the bay ½ mile S of Gaspee Point. 1924.

3 feet wind-blown sand.
8 feet sand and, at the top, gravel.
2 feet silty varved clay with ripple marks.
7 inches silty clay with ripple marks, varves 87–102.
2½ feet yellow-white, on the whole distinctly varved, clay silt, varves 1–86. Winter layers mostly thin.
Sea level. Unknown depth to substratum.

Series measured: Not connected with any other measurement. 1–102.

142.—Providence, 7½ miles SSE of the station, clay pit ¾ mile W of Barrington station and 200 yards S of the railroad track. 1924.

7 feet silty sand and gravel.
2 feet varved silt.
4½ feet varved clay silt, 157 varves.
Varved quicksand. The deposit reported to be 60 feet thick.

Series measured: Not connected with any other. 1–151.

143.—Providence, 2 miles E of the station, 100 yards NE of the northern highway bridge across the Seekonk River, railroad cut between the diverging tracks. 1924.

12 feet fine sand.
6 inches clay silt.
2¾ feet yellow-white, varved clay silt. Winter layers thin and sometimes difficult to distinguish, although the lamination is very marked. About 54 varves.
2 feet clay silt: thick disturbed varves.
Sandy and gravelly material with plenty of cobbles and scattered boulders—probably till.

Series measured: Not connected with any other. 1–54.

LOCALITIES IN SOUTHEASTERN MASSACHUSETTS

These are shown on Figure 12, p. 118. Plate IV.

144.—Taunton, brickyard 1½ miles SE of the station, just SE of the river and near the railroad. 1924.

2½ feet sand.
2 feet clayey silt.
9½ feet silt with clay layers, 41 varves. Silt oxidized yellow down to varve 29, below which it is gray.
Depth to substratum unknown, but probably not great.

Series measured: Not connected with any other series. 1–41, 1–18.

145.—Middleboro, 4½ miles N of the station, brickyard ¾ mile SSW of Titicut station. 1924.

3–5 feet gravelly sand.
3 feet massive clay silt.
4 feet clay silt, 10 varves.
2½ feet disturbed clay.
6 feet silty, distinctly varved clay, varves 1–89.
Depth to bottom unknown, but certainly not great.

Series measured: Series not connected with any other. 1–72, 16–77, 22–88, 25–81, 45–89.

146.—Bridgewater, brickyard 500 yards NE of the station. 1924. Profile composed from several measurements at various places in the large clay pit.

3 feet gravelly sand.
3 feet clayey and sandy silt.
1⅓ feet silty sand, varves 147–165. Sharply set off against the underlying clay.
2 feet stiff clay, varves 89–146. Summer layers of white silt; winter layers dark gray and thick. Sharply set off against the underlying clay.
14 inches lean clay, varves 57–88.
7½ feet distinctly varved clay, silty at the bottom, but growing fairly stiff upwards, varves 1–56.
Fault. Some varves probably missing.
4⅓ feet silty clay, 12 varves.
1 foot disturbed silty clay.
11 inches silty clay, 15 varves.
Sand at bottom of clay pit. Depth to substratum unknown.

Series measured: Series not connected with any other. 1–50, 7–58, 11–74, 11–61, 25–120, 39–94, 80–116, 86–126, 100–151, 145–165.

147.—Bridgewater, 3 miles N of station, brickyard ½ mile N of Elmwood station. 1924.

3 feet clay, sand, and fine gravel in mixture.
7½ feet clay silt, 19 varves.
Not far to substratum.

Series measured: Series not connected with any other. 1–19, 4–19.

### Localities in Vermont

These are shown on Figure 13, p. 120 and Figures 14–16; also Figure 2, p. 95. Plates V and VI.

148.—1½ miles W of Montpelier, railroad cut 200 yards S of Montpelier Junction station house. 1922.

7 feet sand and silt.
3 feet contorted clay.

148.—*Continued*

Distinctly varved brown clay with three thin disturbed zones each representing a few varves, varves 7066–7282.
3 feet disturbed clay.
20 feet till exposed to the level of the track.

Series measured: 7066–7282.

149.—3 miles WNW of Montpelier, slide 300 yards S of the junction of Jones Brook with Winooski River. 1922.

Surface of sedimentation.
2½ feet gravel with cobbles.
13 feet silt and quicksand. Varves not measurable.
16 feet covered.
7 feet contorted clay silt.
Distinctly varved clay, varves 7059–7288.
15 feet varved silt, more or less covered. Depth to substratum unknown.

Series measured 7059–7288.

150.—Brickyard ⅓ mile SE of Waterbury, just E of the railroad, 1921.

3 feet somewhat gravelly sand.
7 feet gravel.
16 feet silt.
20 feet rather stiff clay in thick varves, which are difficult to distinguish; not measured.
Clay with varves containing some silt in the middle part and otherwise consisting of clay that becomes more and more fine-grained upward. Much of the material must have come from the Lamoille Valley. The thicknesses of the varves in order of deposition are in centimeters: 36, 26, 13, 11, 21, 14, 12, 11, 12.5, 11, 16, 17, 16, 17, 9, 9; slidden zone 0.5 m., probably 2 varves; 45, 32, 23, 15, 18, 26, 25, 21, 18.
7 feet contorted stiff clay, perhaps a few thick varves. Coalescing with the lake in the Lamoille Valley via valley at Stowe. (Cf. p. 94.)
Thick bed of distinctly varved clay with a few contorted zones, each representing a single varve to a few varves. Average thickness of varves about 1¾ inches. Varves 7067–7292.

150.—*Continued*

1 ½ feet contorted clay.
7 feet varved silt, 11 varves.
15 feet contorted varved silt. Depth to substratum unknown

Series measured: 7067–7292, and parts not connected with other series.

151.—3⅓ miles NNE of Waterbury, ⅓ mile W of Waterbury Center, slide on Bryant Brook, just W of the highway. 1922.

20 feet covered.
3 feet varved sandy silt, 9 varves.
8 feet disturbed dark brown silty clay in thick varves. Drainage of Lake Lamoille into Lake Winooski.
Gray-brown, beautifully varved clay with two disturbed zones representing 2 and 7 varves respectively, varves 7066–7993.
2 ½ feet thoroughly contorted sandy silt.
4 feet covered to brook level. Depth to bottom unknown.

Series measured: 7066–7293.

152.—3½ miles N of Waterbury, slide just N of Waterbury River and W of the highway bridge across the river. 1922.

45 feet clay, silt, and sand more or less covered and partly disturbed.
3 feet distinctly varved clay, varves 7057–7076.
Thick bed of clay, mostly covered and partly contorted.

Series measured: 7057–7076.

153.—2¾ miles ESE of Essex Junction, slide 200 yards E of Alder Brook and 200 yards N of Winooski River. 1922.

2 feet sand.
2 feet gravel.
Distinctly varved clay. Varves at an average 1 ½ inches thick and consisting of gray-blue silt and red-brown clay. Varves Essex 56–284. Varve 156 consists of quicksilt and is 2 feet thick. It marks a drainage in the vicinity into the Winooski Valley.
Covered many feet to substratum.

Series measured: Essex 56–284.

154.—Essex Junction, brickyard on Indian Brook ½ mile E of the station. 1921.

6 feet clay, not measurable, partly disturbed.
2 feet clay with a thin disturbed zone. Probably varves Essex 69–87, 97–110.
3 feet contorted varved clay.
Till.

Series measured: (Probably Essex 69–87, 97–110).
At other places in the long exposure the greatly contorted clay contains pebbles that may have been kneaded in during advance of the ice border.

155.—Essex Center, slide on Alder Brook ⅔ mile WSW of the village. 1922.

10 feet silt and sand.
10 feet varved sandy silt, winter layers indistinct.
1 ½ feet slidden silt.
Thick bed of beautifully varved clay, varves Essex 17–309. The series shows excellent agreement with the other measurements but has not been used in calculating the normal curve on account of the thickness of the varves.
Brook level. Not far to substratum.

Series measured: (Essex 17–309).

156.—Essex Center, slide on Alder Brook 200 yards SW of the southern highway bridge. 1922.

8 feet sand.
8 feet indistinctly varved silty sand.
3 feet contorted silt.
Thick bed of distinctly varved clay with two slidden zones 1 foot and 1 ½ feet thick respectively. Varves Essex 10-132, 150–171, 196–302.
Brook level. A few feet to bottom.

Series measured: Essex 10–132, 150–171, 196–302.

157.—Slide close by section 156. 1922.

Surface of sedimentation.
8 feet silt and sand.

157.—*Continued*

13 feet quicksilt, varves not measurable.
Quicksilt, varves 303–345.
1¾ inches stiff brown clay, varve 302, indicating large drainage far away.
Distinctly varved clay that grows lean towards the top, varves 38–301.
5 feet covered to brook level. Depth to bottom not great.

Series measured: Essex 38–344, 77–175, 305–345.

158.—Slide 100 yards NE of profile 156. 1922.

Plain of sedimentation.
35 feet coarse silt and sand.
1 foot silty clay.
Clay, varves 197–220.
6 inches disturbed zone, varves 105–196 missing.
Silty clay, varves 57–104.
Quicksand, varves 26–57.
Rather stiff clay, varves 1–25.
5 feet till to level of brook.

Series measured: Essex 1–104, 1–19, 197–220.

159.—2½ miles E of Essex Center, ⅔ mile W of Jericho village, slide on the brook 200 yards N of the highway fork. 1922.

Plain of sedimentation.
2 feet sand and gravel.
8 feet quicksand.
Thinly varved clay, varves 249–279.
1 foot slidden zone.
Clay, varves 203-238.
10 feet sand and silt, disturbed, partly covered.
3 feet silty clay, 21 varves.
10 feet thickly varved silty sand, mostly covered.
2½ feet stiff blue clay, 7 varves.
Brook level. Depth to bottom unknown.

Series measured: Essex 203–238, 249–279, and parts not connected.

160.—8½ miles NNE of Burlington, slide on Malletts Brook at the line between the towns of Colchester and Milton. 1922.

(The clay near the profile is covered by gravel and sand, more than 15 feet thick.)
Surface of erosion.
2 feet massive clay.
Extremely stiff clay at the bottom blue, at the top red, and for the most part brown. Varves partly difficult to distinguish. 175 varves.
Massive, very stiff clay. At least 7 feet to substratum.

Series measured: 1–175, not connected with any other series.

161.—16 miles NE of Burlington, 1¼ miles S of Fairfax, bluff on the east side of Browns River, 200 yards SE of the highway bridge. 1925.

75 to 100 feet of sand.
12 feet sand and silt.
1 foot fine sand, drainage varve.
11 feet clayey silt, about 90 varves of greatly varying thicknesses.
4 inches slidden zone.
1¼ feet silty clay, 25 varves.
2 inches disturbed zone.
1¾ feet clayey silt, varves 139–158.
1½ feet stiff, lead-colored clay with thin shiny winter layers, varves 91–138.
2 inches slidden zone, varve 90.
3¼ feet fat, lead-colored clay, varves 23–89.
3 feet covered to river level. Depth to substratum unknown.

Series measured: Fairfax 23–89, 91–158.

162.—450 yards NW of section 161, bluff at the sharp bend of the river. 1925.

Covered.
5 feet stiff lead-colored clay, varves 16–101.
3 feet covered to river level. Depth to bottom not known.

Series measured: Fairfax 16–101.

163.—⅔ mile NNE of Fairfax, bluff on Mill Brook. 1922.

Plain of sedimentation.
25 feet sand and silt.
2 feet quicksand.
Silty clay, distinctly separated from the underlying stiff clay, varves 139–178.
Stiff gray-brown clay, varve limits partly difficult to distinguish, varves 1–138.
5 inches disturbed zone.
11 feet partly disturbed varved clay, some 70 varves.
3 feet till to brook level.

Series measured: Fairfax 1–178.

164.—17 miles NE of Burlington, 4 miles SW of Cambridge, bluff on Beaver Brook 1 mile N of Cloverdale station and 300 yards W of the railroad track. Two sections. 1922.

A.—

3 feet disturbed clay.
Distinctly varved clay, summer layers of gray silt, winter layers of gray-brown clay, varves 60–116. The thicknesses of the varves in centimeters, in order of deposition are: 18, 8.5, 7, 7.5, 5, 9, 6.5, 8, 3.5, 2, 6.5, 5.5, 4.5, 4, 6, 7, 4.5, 9, 6, 12.5, 4, 4.5, 5, 6, 6.5, 18, 5, 9, 4, 6.5, 4, 5, 5, 8, 8, 5, 4, 5.5, 4, 4.5, 4, 3, 3, 4, 3, 3, 2.7, 3, 2.8, 3, 2, 2, 3.5, 2, 3.5, 8, 2.5.
Brook level. Depth to bottom unknown.

Series measured: Essex 60–116.

B.—Close to A.

1 foot gravel.
1½ feet clayey silt. Mostly thin, not measurable varves.
1½ feet fine sand, 3 to 4 drainage varves.
7 inches stiff clay, probably 3 varves.
Exceedingly greasy clay with thin streaks of fine sand, blue, towards the top brown. Varves partly difficult to distinguish. Not surely correlated, but probably varves Essex 120–303.
13 feet covered to brook level.

Series measured: (Probably Essex 120–303).

165.—2½ miles W of Cambridge, 500 yards W of the sharp bend of the railroad, slide on the highway following the southern bank of Lamoille River. 1922.

Covered.
Distinctly varved clay, varves Essex 145–165.
5 feet covered to the level of the road. Depth to substratum probably great.

Series measured: (Essex 145–165).

166.—2 miles W of Cambridge, bluff on the northern side of Lamoille River, where it turns towards the south to come close to the sharp bend of the railroad. 1922.

Surface of erosion.
10 feet silty sand.
8 feet silty sand, varves Essex 303–316.
2 inches clay, varve 302.
Gray-brown clay for the most part distinctly varved, stiff in the lower part, growing lean upwards, varves 181–301.
Gray-brown silty clay with distinct varvity, varves 131–180.
25 feet covered to bed rock in river.

Series measured: Essex 131–316.

167.—1 mile N of Cambridge, slide on brook just S of the crossroads. 1922.

Surface of erosion.
4 feet covered.
Distinctly varved silt and sand. The thicknesses of the varves in centimeters are in order of deposition: 75, 30, 50, 70, 47, 37, 29, 27, 27, 21, 22, 15, 12.5, 13.5, 17, 11, 13, 14, 10, 8, 9.5, 9, 9, 12, 8.5, and 7.5. The series is not connected with any other.
5 feet covered.
Till.

168.—Cambridge, slide on the northern side of Lamoille River, 200 yards E of the bridge across the river at the eastern edge of the village. 1922.

25 feet clay, silt, sand, and gravel, mostly covered. The clay slidden.
Stiff blue clay, varves Essex 181–237.

168.—*Continued*

7 feet coarse silt and quicksand. Thicknesses of varves mostly worthless, varves 142–180.
Distinctly varved clay silt, varves 72–141.
30 feet covered to bed rock in river.

Series measured: Essex 78–133, 180–237, (134–179).

169.—Jeffersonville, slide on brook at eastern edge of village, 125 yards S of the highway to Cambridge Junction. 1922.

7 feet covered.
Gray-brown distinctly varved clay silt with a contorted zone 6 inches thick. The thicknesses of the varves in centimeters, in order of deposition, are: 11, 10.5, 15.5, 15, 9.5, 11, 11, 23, 9.5, 9, 7, 10.5, 14, 9, 8, 10, 9, 9, 15, 8, 7, 10, 13, 14, 11.5, 9, 11, 11, 15, 15, 15.5; 15 cm. contorted clay silt; 16, 19, 10, 6, 10.5, 8, 7, 5, 7, 8, 9, 12, 7, 8.5, 7.5, 24, 6.5, and 7. Varve number 1 is the same as number 3 at locality 170. Probably varves Essex 91–121, 123–139.
30 feet covered to brook, whose bed is cut in till.

Series measured: (Probably Essex 91–121, 123–139).

170.—300 yards S of profile 169, slide on the brook. 1922.

10 feet covered.
8 feet unvarved sand.
Varved sand. The varves, in order of deposition are 28, 45, 28, 24, 40, and 50 cm. thick.
8 inches slidden zone.
2 ½ feet quicksand, 1 varve.
Varved sandy silt. The thicknesses of the varves in centimeters in order of deposition, are: 21, 16, 14, 16, 14, 20, 15, 14.5, 13.5, 9, 15.5, 17.5, 14, 11, 13.5, 14, 11.5, 17, 9, 10, 11.5, 12.5, 21, 13, 11, 14, 14, 16, 17, 17.5, 27, 21, 20, 10, 16, 13.5, 11.5, 8, 12, 9.5, 11, 22, 12, 12, 7, 19, 8.5, 8, 11, and 14. Varve number 3 is the same as varve number 1 at section 169. Probably varves Essex 93–149.
35 feet covered to brook level. In the brook, till.

Series measured: (Probably Essex 93–149).

171.—St. Johnsbury, at the junction of the railroads at the southern edge of the city, kettle in the southern end of the large gravel deposit. 1921.

171.—*Continued*

45 feet of alternating beds of varved clay, more or less disturbed, and of gravel. The varves at an average 1 inch thick. The deposit contains at least 200 varves.
30 feet coarse glacial gravel exposed.

172.—1 mile W of St. Johnsbury, bluff on the brook at the railroad. 1921.

3 feet leached clay.
Thick bed of lead-colored indistinctly varved clay, very stiff but with lenses and layers of sand. Varves at an average 1¼ inches thick. A few contorted zones. Series probably representing about varves 7250–7520. Depth to bottom unknown.

Series measured: (Probably 7250–7520 with a few gaps).

173.—1 mile W of St. Johnsbury, slide 250 yards WNW of the highway bridge across the river. 1921.

10 feet covered.
13 feet very thinly varved clay, hardly measurable.
Thick bed of lead-colored clay, extremely greasy except for numerous layers of sand, several of which a few inches thick, so that the varves are of greatly varying thickness. Varve limits frequently difficult to distinguish. Measured 280 varves. Series not correlated with any other; in fact it is not known whether it comes below or above that at locality 172.
More than 15 feet to substratum.

174.—1 mile N of St. Johnsbury, 400 yards N of the highway bridge across the Passumpsic River, bluff on the brook E of the highway. Two sections. 1921.

A.—150 yards E of the highway.

15 feet covered.
6 feet varved silt, not measured.
Clay, varves at an average ⅓ inch thick, not very characteristic. Probably varves 7610–7800.
At least 40 feet to substratum.

Series measured: (Probably 7610–7800).

B.—250 yards E of the highway.

30 feet silt and sand.
10 feet silty and sandy clay. Varve surfaces wavy on account of shallow water.
Thick bed of varved clay, for the most part exceedingly greasy, but with occasional layers of sand. Varve limits frequently indistinct. Varves fairly thin, not characteristic. A few contorted zones. Series may represent 7340–7780 with a few gaps.
Depth to substratum unknown.

Series measured: (Probably 7340–7789 with gaps).

175.—9½ miles S of Newport, ½ mile SSE of Orleans station, slide on the edge of the wood. 1921.

7 feet sand.
1 foot leached clay.
1½ feet thinly varved clay, disturbed.
7 inches sandy silt, drainage varve.
Distinctly varved clay silt, 103 varves (Pls. V, VI).
30 feet covered to till.

Series measured: Orleans 1–104, 10–42.

176.—7 miles S of Newport, halfway between Orleans and Coventry, slide at a bend of Barton River 250 yards E of the railroad. 1921.

Clay silt in thick distinct varves. The thicknesses in centimeters of the varves in order of deposition are: 16, 15, 14, 16, 15.5, 13, 12, 10.5, 13, 9, 10, 10.5, 9, 8, 11.5, 8.5, 8, 8.5, 10, 8.5, 11, 8.5, 7, 7, 9, 6, 7.5.
10 feet covered to river level. Depth to substratum unknown.

Series measured: 27 varves, not connected with any other series.

177.—5 miles S of Newport, ½ mile ESE of Coventry station. 1921.

8 feet disturbed and partly covered clay.
Distinctly varved clay. Varves consisting of cream-colored silt and red-brown clay. Somewhat below the top, a

177.—*Continued*

slidden zone; instead of varves missing 20 varves are duplicated. Varves 7–110.

Depth to substratum unknown.

Series measured: Coventry 25–58, 25–44, 50–79, 65–98, 77–110, (7–24, 59–69).

178.—4 miles S of Newport, ½ mile NE of Coventry station, bluff on the north side of the river. 1921.

8 feet clay, sand, and gravel.

Distinctly varved silty clay, with a slidden zone representing 3 varves, varves 1–77.

13 feet covered to river level. Depth to substratum unknown.

Series measured: Coventry 1–31, 11–28, 35–77, 42–56.

179.—2½ miles SSE of Newport, bluff on Cobb Brook 300 yards E of the highway. 1921. (See also Antevs, 1925c.)

Erosion surface.

3 feet till in primary position.

1 foot varved clay, somewhat disturbed, about 10 varves.

3½ feet oxidized, unleached clay. Varves consisting of layers of cream-colored clay silt and of gray-brown clay. Varves Coventry 26–46.

1½ feet till.

10 feet brown varved clay. The upper 9 feet disturbed and faulted. The lowest 14 inches contain 12 partly contorted varves. The entire bed, but especially the 3 topmost feet, strongly weathered.

20 feet till to brook level.

The topmost till bed may mark an oscillation of the ice front and together with the underlying clay and till may date from the disappearance of the last ice sheet. Between the deposition of the lower clay and the second till there must have been a time interval of thousands of years during which the clay was exposed to the atmosphere. Though the locality lies in a belt of slow retreat and repeated readvances of the ice border, the oscillations may not have been long enough to permit so deep

weathering; but the clay bed may have been deposited off one of the Pleistocene ice sheets preceding the Wisconsin ice, and the leaching may have taken place during an interglacial epoch.

180.—3 miles NNW of Newport, bluff on Lake Memphremagog, ⅓ mile W of the lighthouse. 1921.

Erosion surface.
10 feet disturbed clay.
Distinctly varved clay, varves 26–118. Clay mostly contorted in the long section exposed.
1 inch slidden clay.
1½ feet silty sand, disturbed drainage varves.
Lake level. Depth to substratum unknown.

Series measured: Newport 26–118, 26–74, 34–63, 69–118.

181.—3½ miles NNE of Newport, ½ mile S of Lake Park station, bluff on Lake Memphremagog close to the railroad. 1921.

6 feet sand.
Stiff, distinctly varved clay, varves 26–140.
1½ feet contorted zone representing varves 22-25.
Clayey silt, varves 13–21.
Sandy silt, varves 1–20. Varve 1 is bottom varve.
Till at lake level.

Series measured: Newport 1–21, 7–21, 26–140, 26–50, 66–105.

## Localities in Northern Ontario and Quebec

The clay sections studied in Canada as already stated, have been numbered independently from those in the United States. Localities 1–104 were described in a previous paper (Antevs, 1925b, pp. 95–119).

The varves belong to the series Timiskaming, whose varve No. 1 is the bottom varve at locality 63, 79° 27′ 30″ W., 47° 8′ N., at the mouth of Montreal River in Lake Timiskaming (Antevs, 1925b, p. 126). For locations, see Figure 23, p. 142 and Figures 24–27; also Figure 3, p. 98. Plates VII and VIII.

104.—79° 13′ 30″ W., 48° 50′ 30″ N., 3 miles NNW of La Sarre, slide on the west side of La Sarre River at the southernmost rapid (Antevs, 1925b, p. 119; Pls. 7, 8).

At this locality, in 1924, varves 1211–1304 and 1485–1585 were re-measured. The new measurement agrees well with that of 1923.

105.—78° 27′ W., 48° 17′ 15″ N., Kewagama Lake, point on the west side of the peninsula just N of the narrow strait to the eastern part of the lake. 1924.

3 feet disturbed and leached clay.
4 feet clayey silt, 35 varves. Thicknesses of varves in centimeters on order of deposition: 14, 6, 5.5, 7, 10, 9.5, 9, 7, 6, 4, 2.5, 5, 4, 5, 2, 2, 2.5, 1.5, 2.5, 3, 3, 1, 3, 3, 1.7, 1.8, 1.8, 0.7, 1.3, 1, 1.8, 1.3, 1.7, 1.2, 2.3.
2 ½ feet silt, one varve.
20 feet covered to bed rock at the level of the lake.

Series measured: 36 varves, not connected with the normal series, because not characteristic.

106.—¾ mile NE of locality 105, bluff on the east side of the peninsula. 1924.

Surface of sedimentation.
4 inches silt.
2 ½ feet leached clay.
10 feet beautifully varved clay. Summer layers consist of white silt and winter layers of gray-brown clay.
Lake level. Unknown depth to substratum.

Series measured: 919–1141, (919–959, 1066–1092).

107.—½ mile N of locality 106. 1924.

3 feet leached clay.
7 feet distinctly varved clay.
15 feet covered to lake level.

Series measured: 944–1118.

108.—Bluff ¾ mile N of locality 106. 1924.

3 feet weathered clay.
11 feet well varved silty, in the lower part sandy, clay.
Lake level. Probably not far to bottom.

Series measured: 873–1041.

109.—78° 19′ W., 48° 30′ N., 10 miles S of Villemontel station, slide on the east side of Villemontel River. 1924.

7 feet leached and covered clay.
9 feet rather stiff, well varved clay.
River level.

Series measured: 1034–1207.

110.—4 miles S of Villemontel station, slide on the east side of Villemontel River where it bends to flow straight west 1 mile E of the bridge. 1924.

4 feet leached clay.
8 feet well varved clay with a slidden zone of 5 inches representing one varve.
River level. Depth to bottom unknown.

Series measured: 1023–1126 (1072–1100).

111.—400 yards W of locality 110, slide on the north side of the river. 1924.

4 feet leached clay.
11 feet distinctly varved clay. The clay locally disturbed.
5 feet covered to river level.

Series measured: 1075–1310, 1201–1241, (1094–1133, 1142–1200).

112.—79° 22′ 20″ W., 48° 14′ 30″ N., bluff on Dasserat Lake, 200 yards N of the strait to Mishikwish Lake. 1924.

3 feet leached clay.
11 feet well varved clay, gray in the lower part, but growing brown upward.
3 feet below lake level. Depth to bottom unknown.

Series measured: 914–1109, 1069–1087.

113.—1⅛ miles WSW of locality 112, bluff on the west side of the little narrow peninsula. 1924.

3 feet leached clay.
3½ feet stiff clay with a few disturbed zones. About 110 varves, beginning with varve 1201.
9 feet gray-brown well varved clay with a disturbed zone of 8 inches representing 6 varves.
Lake level.

Series measured: 990–1076, 1006–1079, 1083–1200, 1109–1137, 1152–1176.

114.—79° 27′ 15″ W., 48° 17′ N., bluff on the north shore of Dasserat Lake. 1924.

2½ feet leached and disturbed clay.
4 feet stiff, brown clay, varves 1067–1196.
2 inches slidden zone.
6 feet distinctly varved, in the lower part silty, clay; varves 954–1043.
Lake level. Depth to bottom unknown.

Series measured: 954–1043, 954–973, 996–1043, 1067–1196, 1067–1109, 1070–1101, 1116–1143, 1164–1196, (1018–1043).

115.—79° 14′ 30″ W., 48° 30′ N., bluff on the north shore of Duparquet Lake. 1924.

4 feet weathered clay.
6 feet gray-brown clay, varves 1279–1450.
Fault.
4 feet gray-brown fairly stiff clay, varves 1133–1218.
Covered many feet to substratum.

Series measured: 1279–1450, 1318–1352, 1364–1390, 1401–1423, (1133–1218).

116.—79° 17′ 30″ W., 48° 36′ N., slide on the west side of the Abitibi River. 1924.

5 feet weathered clay.
10 feet beautifully varved clay, growing very stiff towards the top. A slidden zone of 4 inches represents 2 varves.
3 feet below river level. Depth to substratum unknown.

Series measured: 1218–1261, 1218–1261, 1264–1420.

117.—79° 33′ 30″ W., 48° 36′ N., ¾ mile N of the mouth of the Madawanasaga River in Lake Abitibi, bluff on point at the Indian cabin. 1924.

3 feet leached clay.
8 feet beautifully varved clay.
4 feet clay partly disturbed and covered to lake level. Depth to bottom not known.

Series measured: 1201–1275, (1169–1200, 1171–1206).

118.—79° 9′ W., 48° 52′ N., 4 miles NNE of La Sarre, bank on the La Sarre River at the power station. 1924.

3 feet weathered clay.
10½ feet brown clay becoming very stiff a little above the bottom. Varve 1208 is bottom varve.
3 feet till.
Bed rock.

Series measured: (1208–1344).

119.—80° 4′ W., 48° 19′ N., bluff on Black River ½ mile W of High Falls, at the sharp bend just below the rapids. 1924.

6 feet disturbed and weathered clay.
7 feet well varved clay. Summer layers of light silt; winter layers of gray-brown clay. 51 varves.
10 feet covered to river.

Series measured: 51 varves corresponding to varves 23–73 at locality 122. Agreement excellent.

120.—200 yards N of locality 119, slide on the north side of the river just below the portage. 1924.

13 feet covered.
7½ feet silty clay, 44 varves. Thicknesses of the varves in centimeters in order of deposition are: 17 (?), 12, 4.5, 2.5, 3, 3.5, 2.5, 2.5, 3, 11.5, 5(?), 5.5, 6, 3.5, 6, 5.5(?), 8.5(?), 2.5, 3.5, 2, 5.5(?), 2.5, 2.5, 5.5, 4, 4, 7, 3, 2.5, 5.5, 5.5, 6, 8, 12, 4, 7, 6.5, 7, 6.3, 6, 4, 4.
River level. Probably not deep to bottom.

Series measured: 44 varves not connected with any other series.

121.—Slide on the river 200 yards W of locality 120. 1924.

Several feet covered.
10 feet silty clay with two slidden zones. 75 varves.
3 feet covered to river level.

Series measured: 75 varves, not connected with any other series.

122.—300 yards NW of locality 119, slide on the north side of the river at a sharp bend. 1924.

4 feet covered.
9½ feet clay with distinct varvity. Summer layers of light silt; winter layers of gray-brown clay. 86 varves.
6 feet covered to river.

Series measured: 86 varves connected with the series at locality 119.

123.—80° 45′ W., 48° 50′ N., 5½ miles NW of Iroquois Falls, slide in the eastern bank of Abitibi River. 1925.

Erosion surface.
3 feet leached clay.
5½ feet stiff, gray-brown clay.
5 feet contorted clay to river level. Depth to bottom unknown.

Series measured: (1486–1550). Agreement with other measurements is excellent.

124.—80° 57′ W., 48° 44′ N., 8 miles WNW of Porquis Junction, bluff on Frederick House River at the mouth of Slim Creek. 1924.

1 foot peat and sand.
1 foot clayey silt.
2 inches postglacial sand with shells.
3 feet yellow, unvarved, probably postglacial clay. No shells observed.
7½ inches clay with increasing percentage of silt. Varves 2015–2027.

124.—*Continued*

9 inches stiff, brown clay. Varve limits fairly distinct. Varves 1965–2014.
1 foot stiff clay. Varve limits partly indistinct. Varves 1882–1964.
3 feet brown clay with good varvity, varves 1813–1881.
4 feet covered to river level. Far to substratum.

Series measured: 1813–1881, 1965–2026, (1882–1964).

125.—½ mile NW of locality 124, slide on the east side of the river, opposite the lower end of the abandoned river channel. 1924.

4 feet clay and plant remains.
1 foot yellowish, unvarved, silty clay.
6½ feet very stiff, brown clay, for the most part thinly varved.
2 feet covered to river. Deep to substratum.

Series measured: 1760–2027.

126.—1 mile NW of locality 124, slide on the east bank of the river just above the rapid. 1924.

2 feet massive postglacial clay.
Erosion level.
12 feet stiff, brown well varved clay with a disturbed zone of 4 inches representing 2 varves.
2 feet covered to river. Depth to bottom unknown.

Series measured: 1543–1600, (1486–1542, 1601–1670).

127.—200 yards N of locality 126 and the rapid, slide in the eastern river bank. 1924.

10 feet contorted clay with scattered boulders; not regular boulder clay.
2½ feet stiff, brown clay, varves 1647–1688.
Several feet not measured, partly contorted clay.
9½ feet distinctly varved clay, in the lower part silty, in the upper part greasy, varves 1432–1542.
15 feet contorted varved silt to river level.

Series measured: 1432–1542, 1647–1688.

128.—81° W., 48° 47′ N., slide on the east side of Frederick House River just below the fall. 1924.

Many feet typical boulder clay.
1 foot contorted varved clay.
9 feet greasy clay, varves 1595–1808.
A few feet covered.
4½ feet stiff clay, varves 1487–1547.
5 feet silt in disturbed thick varves.
Till and bed rock to river level.

Series measured: 1485–1547, 1695–1760, (1761–1808).

129.—⅔ mile NW of the fall and locality 128, slide on the west side of the river. 1924.

10 feet typical boulder clay.
Many feet covered.
14½ feet stiff, well varved clay.
10 feet covered to river.

Series measured: 1601–1760, (1503–1600).

130.—1 mile NW of the fall and locality 128, slide on the west side of the river. 1924.

15 feet till.
4 feet stiff clay in thin varves. A slidden zone of 4 inches represents 3 varves.
Covered. Far to bottom.

Series measured: 1823–1964, (1965–1979).

131.—400 yards NW of locality 130, slide in the western river bank. 1924.

30 feet boulder clay.
1⅓ feet stiff clay, about 80 varves.
2 feet contorted clay.
14½ feet well varved clay, varves 1520–1690.
15 feet covered to river.

Series measured: 1601–1690, (1520–1600).

132.—1½ miles NW of the fall and locality 128, 200 yards N of a rapid. 1924.

A.—Slide in the east bank of the river.

Many feet boulder clay.
5 feet contorted clay.
8 feet stiff clay.
15 feet covered to river.

Series measured: 1603–1760, (1761–1784).

B.—Slide in the west bank of the river.

30 feet boulder clay.
5½ feet exceedingly stiff, brown clay with distinct varvity.
15 feet covered to river.

Series measured: 1674–1844.

133.—2 miles NW of locality 128 and the fall, and 300 yards SE of the new river trench formed by diversion of the river in 1909 (see Knight, etc., 1919, p. 42; our p. 171), slide in the eastern river bank. 1924.

30 feet stiff boulder clay.
10 inches stiff clay, about 40 varves.
11 feet greasy clay with a disturbed zone of 4 inches representing 3 varves, varves 1647–1950.
15 feet covered to river.

Series measured: 1647–1842, (1843–1950).

134.—150 yards from locality 133, 150 yards S of the new river channel, slide on the W side of the river. 1924.

30 feet boulder clay.
4½ feet contorted thinly varved clay.
1½ feet exceedingly fat clay, varves 1698–1740.
11 feet not measured.
4½ feet well varved clay, varves 1484–1547.
12 feet partly disturbed varved silt, probably about 70 varves.
10 feet covered to river level. Not deep to bed rock which outcrops in the river 30 yards from the locality.

Series measured: 1484–1547, (1698–1740).

135.—150 yards N of locality 134, the east side of the new river channel. 1924.

23 feet light-gray boulder clay with scattered pebbles.
2½ feet contorted varved clay.
10 inches stiff clay, varves 1695–1719.
8 inches disturbed varved clay.
2½ feet well varved greasy clay, varves 1640–1690.
7 feet disturbed clay and silt.
6 feet beautifully varved clay, varves 1488–1582.
15 feet clay and silt partly disturbed and covered to river level.

Series measured: 1488–1582, 1640–1690, (1695–1719).

136.—81° 9′ W., 49° 3′ N., 5½ miles W of Cochrane, 3 miles S of the railway bridge across the Frederick House River, ½ mile N of the rapid, slide in the east river bank. 1924.

25 feet partly covered, largely boulder clay.
7 feet well varved silty clay, 34 varves.
15 feet covered to river.

Series measured: 34 varves, not connected with any other series.

137.—2½ miles N of locality 136, ⅔ mile S of the railway bridge, just S of the mouth of a little brook, slide in the eastern river bank. 1924.

20 feet mostly covered, till and, below, varved clay.
12 feet clayey, at the base sandy, silt, 60 varves.
12 feet covered to river.

Series measured: 60 varves not connected with any other series.

138.—5½ miles W of Cochrane, 300 yards S of the railway bridge across the Frederick House River, slide on the east side of the river. 1924.

25 feet till consisting of clay, sand, and boulders.
6 feet massive, greasy clay.
4½ feet stiff clay with varves increasingly thick upwards; varves 236–260.

138.—*Continued*

16 feet very stiff well varved clay, varves 26–235.
Fault.
8 feet clay, varves 1–24.
River level. Depth to substratum unknown.

Series measured: 1–24, 26–260, not connected with any other series.

139.—81° 23′ W., 48° 31′ N., 4½ miles NW of Timmins, slide in the north bank of the Mattagami River. 1925.

5 feet leached clay.
3 feet well varved clay, varves 1646–1699.
Fault.
4 feet yellow-brown clay, varves 1568–1620.
River level. Deep to substratum.

Series measured: (1568–1620, 1646–1699). Agreement with other series fairly good.

140.—1¼ miles W of locality 139, slide on the south side of the river. 1925.

7 feet covered and leached clay.
5½ feet well varved clay.
10 feet covered to river.

Series measured: (Probably 1451–1517). Agreement with other measurements not good.

141.—81° 36′ W., 49° 6′ N., 26 miles W of Cochrane, slide on the east side of the Mattagami River just N of the mouth of Loon River. 1925.

25 feet boulder clay.
20 feet covered.
4 feet contorted varved clay.
4 feet well varved clay, 32 varves. Thicknesses of varves in centimeters in order of deposition are: 5, 3.5, 3.5, 5, 2.5, 3, 3.3, 3.5, 4, 3.7, 4.5, 3, 3.5, 3.5, 3.5, 3, 4.7, 3, 3.5, 3 7, 3.3, 3.8, 3.5, 3.5, 3.5, 2.8, 3, 2.8, 3.2, 3, 3.3, 3.2.
5 feet covered to river level. Depth to bottom unknown.

Series measured: 32 varves not connected with any other series.

142.—83° 15′ W., 49° 36′ N., Mattice, road cut at the brook a few hundred yards E of the railway station. Two sections. 1924.

A.—

2 feet weathered clay.

Clay. Thicknesses in centimeters of the varves in order of deposition are: 9, 5, 4.5, 6, 6, 5, 4.5, 6, 4.5, 3.5, 4, 6, 4.5, 4, 4.5, 4, 3, 4. Some 10–15 feet to substratum.

B.—Close by A.

1 ½ feet leached clay.

Clay. Thicknesses in centimeters of the varves in order of formation are: 10, 11, 10.5, 9.5, 5, 7, 7.5, 5.5, 5, 5, 4, 4, 5.5, 5, 4, 5, 4, 5.

The two series are not correlated with each other or with any other series.

143.—A little W of locality 142 and Mattice station, gully on the west side of Missinaibi River, 150 yards N of the railway bridge. 1924.

2 feet weathered clay, 10–15 varves.
2 feet silty till.
4 feet boulder clay.
4 feet sandy and silty till.
2 feet contorted varved clay with boulders kneaded into it.
8 inches stiff, dark-gray clay, varves 22–47.
1 foot lean clay, varves 1–21.
3 feet boulder clay.

Series measured: 1–47, 1–30, 1–43, not connected with any other series.

### Localities in Southern Quebec and Ontario

Most of these are shown on Figure 16, p. 126, Figure 21, p. 136, and Figure 22, p. 139; also Figure 4, p. 99 and Figure 29, p. 164. Plates VI and VII.

144.—72° 7′ W., 45° N., 30 miles SSW of Sherbrooke, just N of the International Boundary, 1½ miles E of Beebe Junction, bluff on the north side of Tomifobia River. 1923.

30 feet sand.
15 feet distinctly varved clay with three slidden zones representing unknown numbers of varves. Varves 1–272. Varves 31, 32, and 107 contain sand and record drainages.
20 feet till exposed to river level.

Series measured: The series shows no correspondence to that on Lake Memphremagog 4 miles to the southwest and is not connected with any other series; 1–132, 44–100, 138–149, 152–204, 208–272.

145.—2 miles NNW of locality 144, bluff in the eastern bank of Tomifobia River. 1923.

4 feet gravel.
(Disconformity.)
Thick bed of distinctly varved clay slit with a slidden zone 1½ inches thick. Summer layers consisting of cream-colored fine silt and winter layers of stiff brownish clay. Varve 5 contains sand and marks a drainage in the vicinity.
Clay extends to unknown depth below river level.

Series measured: Not correlated with any other varve series; 1–60, 14–57.

146.—4½ miles N of locality 144, 1½ miles NE of Tomifobia station, slide on the west bank of Tomifobia River, 100 yards W of the second railroad bridge across the river. 1923.

3 feet gravel.
Silty clay, varves 29–51, registering a remarkable increase in the drainage area.
Greasy, yellow clay, varves 1–28.
Faults. 30 feet covered to river level. Depth to bottom unknown.

Series measured: Connected only with locality 147; 1–51.

147.—½ mile NE of locality 146, bluff on the east side of the river, 200 yards E of the railroad track. 1923.

Covered.

Varved silty sand, varves 29–46. Increase in deposition still much more marked than at locality 146. The thicknesses in centimeters of the varves in order of deposition are: 56, 20, 13, 11, 7, 8, 13, 5, 14, 11, 6, 8, 7, 8.5, 12, 16, 7, 6.

Clay, varves 24–28. Thicknesses: 3, 2.5, 3, 2.5, 2 cm.

2 feet disturbed zone, representing many varves.

Blue varved silt, not correlated with locality 146. The thicknesses in centimeters of the varves are: 16, 8, 28, 7.5, 8, 12, 10, 7, 12, 9, 8, 6, 4.5, 5, 8, 7.

15 feet covered to river level.

Series measured: 24–46 correlated with locality 146, and 16 varves not correlated with any other series.

148.—71° 46′ W., 44° 28′ N., 7 miles NE of Sherbrooke, clay pit 300 yards NW of Ascot station. 1923.

1½ feet clayey gravel, perhaps till.

1½ feet weathered clay.

2½ feet clay. The thicknesses of the varves in centimeters are in order of deposition: 4.5, 10.5, 5.5, 7, 3.5, 6, 3.5, 3, 5.5, 5, 5, 6.5, 6, 5.5, 9.

Till.

Series measured: Not connected with any other; 15 varves.

149.—72° 9′ W., 45° 40′ N., 23 miles NW of Sherbrooke, brickyard 1 mile N of Richmond station. 1923.

4½ feet contorted clay silt.

1½ feet clay, 55 varves. Limits of varves 1–17 are indistinct. Varve 30 contains coarse silt.

5 feet disturbed clay.

Till.

Series measured: Not connected with any other series; 1–55, 19–36.

150.—76° 33′ W., 44° 13′ N., 3½ miles W of Kingston, bluff on the west side of the bay into which Little Cataraqui River empties. Two sections. 1925.

150.—*Continued*

A.—

4 feet leached clay.
1½ feet yellow, hard-packed clay with indistinct varvity, varves 56–91.
1 foot contorted clay, representing an unknown number of varves.
3½ feet clay like that above the disturbed zone, varves 1–52.
Level of Lake Ontario. Depth to substratum unknown.

Series measured: Not connected with any other; 1–52, 56–91.

B.—Close by A.

6 feet leached and contorted clay.
3 feet yellow, indistinctly varved clay, 45 varves.
Lake level.

Series measured: Not connected with any series; 1–45.

151.—74° 27′ W., 45° 34′ N., 44 miles W of Montreal, 3 miles W of Point Fortune and 8 miles E of Hawkesbury, bluff on brook 100 yards E of bridge of south-going road, a few hundred yards S of the highway. 1925.

1 foot leached clay.
Stiff clay, three varves, 6, 8.5, and 16 cm. thick respectively.
1 foot brown silty and somewhat gravelly clay.
1 foot brown silty clay.
4 inches indistinctly varved clay.
5 inches stiff gray-brown clay with indistinct varve limits, varves 98–122.
1⅔ feet stiff gray-brown clay with fairly distinct varvity, varves 17–97.
2 feet varved silt, varves 6–16. Varves thick because of proximity of the ice edge.
1½ feet disturbed zone.
Silt, 2 varves, 17 and 22 cm. thick respectively.
4 feet loose, sandy till.
4 feet hard-packed boulder clay exposed to brook level.

Series measured: Correlated only with locality 152; 6–122.

152.—100 yards W of locality 151, road cut at the bridge across the brook. 1925.

7 feet dark-brown, poorly exposed clay.
1½ feet stiff red-brown clay, varves 17–77.
4½ feet silt, varves 1–16.
3–4 feet covered to till in brook, a few varves.

Series measured: Connected only with locality 151; 1–77.

153.—75° 59′ W., 45° 51′ N., 33 miles NNW of Ottawa, 2½ miles SE of Venosta station, slide on brook 250 yards W of the railway. 1924.

6 feet somewhat disturbed clay.
Thick deposit of indistinctly varved clay. Varves consisting of a thin layer of sandy silt and a thick layer of gray-brown clay. The thicknesses in centimeters of the varves in order of deposition are: 12, 15, 12, 13, 14, 19, 14, 25, 14, 10 11, 12, 10, 14, 15, 13, 10, 9.5, 12, 13, 12, 14.5, 16.5, 12, 13, 14, 10, 15, 7, 8, 8, 10.5, 10.5, 12, 9, 9.5, 6.5, 7, 8.5, 9.5, 9.5.
10 feet covered to brook. Depth to bottom unknown.

Series measured: Not correlated with any other; 1–41.

154.—Slide on the brook 150 yards N of locality 153, and 150 yards W of the railway. 1924.

20 feet disturbed and covered.
Indistinctly varved clay. Thicknesses in centimeters of the varves in order of deposition are: 16.5, 8, 11, 10, 15, 13, 10, 10, 7.5, 13.5, 12.5, 12.5, 17, 11.5, 6, 10, 8, 9.
20 feet covered to brook. Depth to substratum not known.

Series measured: Not correlated with any series; 1–18.

155.—76° W., 45° 56′ N., 40 miles NNW of Ottawa, 3 miles ESE of Kazubazua station, road cut some hundred yards S of Kazubazua River and 1½ miles WSW of its junction with the Gatineau River. 1925.

2 feet leached clay.
4½ feet distinctly varved clay, 37 varves. Varves consisting of yellow-brown silt and gray-brown clay. Thick-

155.—*Continued*

nesses of varves in centimeters in order of deposition are: 6.5, 3.3, 3, 4.5, 1.5, 5.5, 4.5, 3.3, 3.5, 3, 5.5, 2.5, 4, 3.5, 4.5, 3.5, 2.5, 3.5, 3.5, 2.5, 3, 2.5, 2.5, 2, 4.5, 4, 6.5, 3.5, 3.5, 3.7, 4.5, 8 (?), 5, 5, 4, 3, 2.5.

A few feet covered to bed rock.

Series measured: Not correlated with any other series; 1–37.

156.—4 miles NNW of locality 155, railway cut 1 mile N of Kazubazua station. 1924.

3 feet sand.

14 feet beautifully varved clay, 69 varves. Varves increase in thickness at top because of shallowing.

Disturbed clay. Depth to bottom unknown.

Series measured: Not correlated with any series; 1–69.

157.—3 miles N of locality 155, Aylwin, slide on the west side of Gatineau River, 300 yards N of the crossroads. 1924.

Alluvial sand.

2 feet disturbed varved clay.

8 feet well-varved clay, 37 varves. The thicknesses in centimeters of the varves in order of deposition are: 8, 7, 6, 7, 4, 7.5, 8, 7.5, 7, 7.5, 8, 7.5, 7.5, 9, 6, 5, 6, 6, 6.5, 8, 7, 7.5, 8, 6.5, 6, 6, 7, 5, 8, 5.5. 6.5, 8, 5.5, 5.5, 7.5, 3.5, 8.

River level. Depth to bottom unknown.

Series measured: Not correlated with any other series; 1–37.

158.—Bluff 100 yards N of locality 157. 1924.

15 feet alluvial sand.

9 feet distinctly varved clay, 59 varves.

River level. Depth to bottom unknown.

Series measured: Not correlated with any other series; 1–59.

159.—75° 57′ W., 46° 26′ N., 73 miles N of Ottawa, 3¾ miles N of Maniwaki, bluff on the west side of Gatineau River, just below Rapide des Eaux. 1924.

3 feet sand and silt.
18 feet light-gray, hard-packed clay silt, 800 varves. Varvity partly good, partly (especially varves 510–530) indistinct.
Probably about 10 feet covered to bed rock.

Series measured: Not connected with any other series; 1–800.

160.—77° 7′ W., 45° 45′ N., cut on the highway 5 miles S of Pembroke. 1925.

10 feet clay and silt.
1⅔ feet clay, varves 131–148.
3 inches disturbed clay representing an unknown number of varves.
6 feet clay, varves 1–127. Varvity fairly distinct in the lower part but becoming faint towards the top.
5 feet sand.
Covered. Depth to bottom unknown.

Series measured: Not correlated with any other series; 1–127, 131–148. The clay may date from the second deep-water stage of the Champlain Sea.

161.—5 miles N of locality 160, brickyard at the southern edge of Pembroke. 1925.

7 feet indistinctly varved clay, about 350 varves.
1½ feet sand.
(Disconformity.)
⅓ foot leached clay.
2 feet clay with fairly distinct varvity, 65 varves.
1 foot, gray-brown, rather stiff, nearly massive clay. Depth to bottom unknown.

Series measured: Not connected with any other series; 1–65. The profile records the two deep-water stages and the intervening shallow-water stage of the Champlain Sea.

162.—79° 25′ W., 45° 58′ N., bluff on the eastern bank of the South River, 3 miles SW of Trout Creek and 1¼ miles SSW of locality 21. 1925.

8 feet disturbed, in the upper part silty, clay.
4 feet well-varved clay, varves 158–243.
1½ feet disturbed zone representing 4 varves.
4 feet distinctly varved clay, varves 72–153.
1 foot disturbed clay.
5 feet covered to river level. Depth to substratum unknown.

Series measured: Connected with locality 21 varves 104–172 but not with other parts of the series (Antevs, 1925b, Pl. 9). The series here published is dated according to this connection, which is based on good agreement. Varves 72–243.

163.—79° 26′ W., 46° 2′ 30″ N., bluff in the eastern bank of South River, 4 miles SW of Powassan. 1925.

8 feet sand.
3 feet contorted, sandy clay.
4 feet gray, sandy clay with indistinct varvity.
4¾ feet clay, 138 varves. Summer layers consisting of gray clayey silt; winter layers of red clay. Varves 4, 5, 10, and 39 might each be two.
1 foot blue, sandy clay with very thin varves.
Level of river. Depth to bottom unknown.

Series measured: Not correlated with any other; 1–138.

164.—A few hundred yards NE of locality 163, on another meander of the river. 1925.

7 feet covered.
3 feet well varved clay, varves 107–151. Plate VII.
5 inches disturbed clay, representing 1 varve.
1⅔ feet clay, varves 88–105.
2 feet covered to river level.

Series measured: Series connected with Powassan, localities 22–25 (Antevs, 1925b, Pl. 5). The agreement is good. The measured varves are 88–105, 107–151.

165.—1¾ miles NE of locality 163, 2½ miles SW of Powassan, slide on the north bank of the South River near the highway. 1925.

1 foot leached clay.
7 feet beautifully varved clay, varves 13–82.
2 feet sandy silt, varves 11 and 12. Drainage which is not recorded at locality 22.
3 inches clay, varves 9 and 10.
1 foot sandy silt, varve No. 8, bottom varve.
5 feet till exposed to river level.

Series measured: Series connected with Powassan, locality 22 (Antevs, 1925b, Pl. 5). Measured, varves 8–82.

LOCALITIES AT TORONTO AND SCARBORO HEIGHTS

These are shown on Figure 17, p. 128, and Figures 18–20. Plate IX.

166.—3 miles NNE of Union station, Toronto, Don Valley Brickyard.

167.—Brickyard ¾ mile NNE of locality 166.

168.—8 miles NE of Union station, Toronto, Scarboro Bluffs on Lake Ontario opposite trolley station 26, western part of Dutch Church.

169.—350 yards NE of section 168.

170.—200 yards NE of profile 169, ravine at the cement foundation, eastern part of Dutch Church.

171.—300 yards NE of Dutch Church.

172.—½ mile NE of Dutch Church.

173.—250 yards SW of section 174.

174.—1 mile NE of Dutch Church, opposite trolley stop 32.

These profiles were measured in 1923 and 1924.

The till beds consist of massive clay with scattered pebbles and occasional boulders. The nature of the varved clays is touched upon under the description of the curves, page 241. All the glacial beds, which are those of particular interest here, have undergone partial consolidation and are very hard, which makes their age uncertain (see p. 129). The beds have been correlated by means of field observations, comparison of the sequence of strata, and by connections of the varve graphs. The correlations, as far as they have been carried out, are indicated in the figures. The principal varve curves, matched with each other, are presented in Plate IX. Profiles 166–170 are completely correlated with each other, and numbers 172–174 practically with each other; but these two groups are not quite satisfactorily connected by means of profile 171.

# CHAPTER XI

## RELIABILITY AND SIGNIFICANCE OF THE VARVES MEASURED

Observation of characteristics of the varved sediments at different localities and horizons, comparison of the graphs made after measurement, and other considerations suggest comments which are made here, in systematic order. Some are of the nature of interpretation of local changes of drainage or elevation during ice recession. Particular attention, however, is directed to the question of the degree of reliability of the different varve curves (Plates I–IX).

These curves, as described in previous papers (Antevs, 1922, p. 47; 1925b, p. 120), are constructed to show as accurately as the material permits the average thickness and number of the annual clay layers during ice retreat. The individual graphs are first matched and corrected for number of varves. If, for example, out of three measurements two agreed, but one had one varve less or more than the others, the exact location of the mistake was determined and the curve corrected by dividing one varve in two or uniting two varves in one, so that this curve agreed with the two others. Then the curves or such parts of them as included undisturbed varves of normal variation and thickness were selected for constructing the *normal curve*, and those curves were discarded that showed great difference in thickness from the majority of poor agreement in the shape of the curve. The normal curve was constructed from the selected individual curves by calculating the average thickness of each single varve. When measurements of the same series were at hand from distant or differently situated parts of the same lake basin or from different regions, separate normal curves were worked out for each region. Varves abnormal at all localities have been marked by broken lines.

### The Peripheral Area

*Hackensack*, N. J., localities 92–99, Plate I.[1]

1–200.—Good. The different measurements agree well. The thicknesses of the varves may, on the whole, faithfully record the ice melting.

1–6 are abnormally thick, as deposited close to the ice edge.

201–400.—Very good up to 300; thereafter good.

366 contains silt and is too thick.

401–600.—Mostly very good. The different measurements show excellent agreement with each other.

478 contains sand and is too thick.

601–670.—Good.

671–952.—Not good. The varve limits are often difficult to distinguish, so that the number is not sure; and the varves are mostly too thin. However, when the varves are correctly measured, the curves show good correspondence with each other.

873–878 have light-red winter layers.

947 begins a series of silty varves with light-red winter layers. It may mark drainage or possibly sudden increase of the drainage area (see p. 112).

In a zone of 20 inches above varve 952 the clay contains a great number of very thin, dark clay layers, and the varve limits are too indistinct to be surely determined. The zone may contain between 75 and 200 varves. In the uppermost part of the zone the clay is fat and gray to red-brown, and the varves are exceedingly thin. The overlying series has been arbitrarily begun with No. 1101.

1101–1200.—Fairly good, although the number of the varves is not quite certain.

1101 begins a series of silty clay with light-red winter layers of similar origin as the series beginning with varve 947.

[1] Varves 20 to 333, 560 to 692, 780 to 852, 1101 to 1120, and 1290 to 1400 are the same as Reeds's (1926) varves −1063 to −750, [118 varves seem to be missing] −641 to −507, −405 to −333, +1 to +20, and 230 to 340.

1201–1490.—Mostly fairly good, but partly not good. The number of the varves somewhat uncertain, and the thicknesses often too small to be characteristic.

*Little Falls*, N. J., locality 101, Plate I.

The series on the whole very good both regarding number and thickness of the varves.

80–86 are pressed.

94, 135, 141, and 171 contain silt and may be abnormally thick.

*New Haven:* Series A and B, localities 103–111, Haverstraw, N. Y. Series C, localities 123–128, New Haven, Conn. Plates I and II.

1–100.—Fairly good. The different measurements, which, however, were made close to each other, agree very well. Varves 1–56, being deposited relatively close to the ice edge, contain more or less sand, so that some of the variations of the thicknesses probably give an exaggerated idea of the variations in the summer heat.

49, 53, 55, and 56, especially, may be abnormally thick.

101–179.—The individual measurements upon which the curves are based show excellent agreement with each other, and so consequently do also the two normal curves.

113, 114, and 125, curve B, are too thick.

131, 132, and 136, both curves, contain some sand and are perhaps too thick.

180–200.—Varve 180 begins the series at New Haven, of which, however, no curve is being given. The individual curves in each group and also the normal curves show striking correspondence. Varves 183, 195, and others, thick both at Haverstraw and New Haven, indicate summers of unusual ice melting.

201–400.—The individual measurements within the three groups agree very well. Curves A and B, i.e. the Haverstraw curves, show practically detailed correspondence with each other and good agreement with curve C, the New

201–400.—*Continued*

Haven curve, so that the connection over the distance of 57 miles (92 km.) is perfectly sure. Because of local conditions the curves, however, present differences in details chiefly in the respect that several varves are too thick at New Haven.

205, 233–235, 250, 286, 332, 354, 392, and several others are unusually thick both at Haverstraw and New Haven and may consequently mark very warm summers.

206, 230, 310, and 396, curve A, are too thick.

206 and 376, curve B, are too thick.

290 and 356, curve B, are too thin.

202, 213, 223, 227, 232, 240, 244, 334, 339, 350, 353, 364, and 365, curve C, are too thick.

401–544.—The individual measurements within the groups correspond in details. Curves A and B agree in details, and these graphs correspond well with curve C, the New Haven curve. Thus the fluctuations carefully record the annual variations in ice melting. Varves 484, 514, and others that are thick both at Haverstraw and New Haven record especially warm summers.

467, 518, and 528, curves A and B, are too thick.

416, 421, 440, 499, 500, and 526, curve C, are too thick.

545–577.—The individual measurements as well as the two curves agree excellently.

578–732.—The different measurements correspond in details, and the curve forms a good record of the variation of the annual rate of the ice melting.

578, 589, 698, and 702–704 are probably too thick.

*Hartford:* Series A, localities 112–121, Newburg, N. Y. Plate III.

Series B, locality 139, Hartford, Conn.

Series C, the curve published as "I Connecticut" in Antevs, 1922, Pl. I. What was then believed to be varves 3017 and 3018 has proved to be one varve, now given number 3018.

*Hartford—Continued*

The numbering of the varves is that given in the paper just referred to (Antevs, 1922, p. 49).[1]

2700–2867.—The curve is exclusively based upon measurements at locality 115, because the actual thicknesses of the varves vary from one locality to another. The fluctuations, however, are very much the same at the other localities, so that the curve presented may form a good record of the annual variations of the ice melting.

2718 may be too thick, and so are perhaps 2794 and 2796.

2868–3000.—The correspondence between the curves is fairly good in spite of the fact that the actual thicknesses of the varves at Newburg are six to seven times as great as at Hartford. Material from the Hudson Valley, not used in the construction of the normal curve, almost corresponds in details with the curve.

2900, 2941, 2953, and 2954, curve A, are too thick.

2933 and 2942, curve B, are too thick.

3000–3121.—Curves B and C show very good correspondence, and these curves agree quite well with curve A, although the varves in the latter graph are about three times as thick as in the former ones and the fluctuations consequently are much greater. The connection between the series from the Hudson Valley and the Connecticut Valley is sure. Several varves are thick both at Newburg and at Hartford and thus may mark particularly warm summers.

3038, 3063, 3105, and 3121, series A, are too thick.

3065, series B, is too thick.

3122–3170.—The two curves correspond fairly well.

*Locality 122.* This locality and the following ones on Plate IV.

1–30.—Varve limits are very distinct, and thicknesses on the whole good.

5, 8, and 13 contain sand and are abnormally thick.

[1] Varves 2826 to 2918 are the same as De Geer's (1926, Pl. 2) varves Dutchess Jc. −5869 to −5778.

31–71.—Varve limits are mostly indistinct, so that the varve number is not entirely certain.

57–61 are abnormally thick and may record a drainage.

*Newfield*, localities 129 and 131.

The different measurements, made close to each other, agree in details.

The upward increase in the thickness of the varves may be due to shallowing of the water or approach of the readvancing ice—the clay was overridden—or to both. Varves 1–28 probably form a good record of the ice melting.

*Locality 132.*

Since the lowest varves were deposited close to the ice edge, their thicknesses are not reliable. Varve 13 is perhaps too thick.

*Berlin*, localities 130, 133–135.

The individual measurements correspond very well, and the lower half of the series may be good. Varves 21–32 are probably too thin, and varves 33–37 are too thick.

*Locality 136.*

Probably good except for varve 2.

*Locality 137.*

The gradual decrease in the thickness of the varves is no doubt due to increasing distance from the ice edge, as it retired.

*Locality 138.*

The lowest varves were deposited too close to the ice edge to be characteristic.

*Locality 141.*

Probably good up to varve 85, above which the layer surfaces are wavy, owing to wave action in shallow water.

*Locality 142.*

The curve may fairly well record the varying rate of the annual ice melting. The increase of the varve thicknesses beginning at varve 106 was perhaps due to enlargement of the drainage area.

*Locality 143.*

On the whole fairly good. Varve 8 may be two varves and varve 18 may be three.

*Locality 144.*

Varves 1 and 14 are probably too thick, and so are most of the varves 28–36, which were deposited in shallow water.

*Locality 145.*

Varves 2, 5, and 6 are abnormally thick, and varves 8–14 are somewhat disturbed; otherwise the curve appears to be good. The individual measurements agree excellently.

*Locality 146.*

Varves 1–18 are evidently too thick because deposited close to the ice edge. Varves 19–151, with few exceptions, may record the ice melting satisfactorily; the individual measurements agree very well. Varves 151–165 were apparently laid down in shallow water. The abrupt transition from lean to stiff clay at 88/89 and again to lean clay at 146/147 and the less pronounced transition from rather fat to lean clay at 56/57 were perhaps principally due to change in the material.

*Locality 147.*

The clay was deposited both in shallow water and close to the ice edge, so that the curve is probably not characteristic.

## Northern New England

*Connecticut Valley*, "22 Vt.," "23 Vt.," localities 85 and 88, Plate V.

The numbering of the varves is that applied in 1922 to the New England clays, according to which the lowest varve measured at Hartford, Conn., has number 3001 (Antevs, 1922).

Measurements made in 1925 show that the varves 7381 and 7391 each are two varves, so that varve now given number 7401 is the same as the varve that has had number 7399.

*Connecticut Valley—Continued*

7401–7461.—The varve measurements at localities 85 and 88 agree well. The clay is exceedingly stiff. The abrupt increase of the varves to more than twice their previous thickness beginning with varve 7401 is remarkable, as the varves, even the lowest ones, consist of exceptionally fine material and lack silt and sand altogether. Nevertheless it may be due, at least to some extent, to a sudden increase of the drainage area, since the abruptness cannot be explained otherwise. The glacial rivers apparently had access to an abundance of very fine-grained material, and flocculation and deposition of the fine mud took place before long transportation. The augmented deposition was also perhaps connected with increased ice melting.

7410 might be two varves.

7446 and 7447 should perhaps be one varve.

7462–7500.—Good. Clay moderately stiff.

7501–7600.—The two measurements made in 1921 and 1925, respectively, agree well.

7506, 7507, 7515, and 7537 may record drainages and be abnormally thick.

7579 and 7580 should perhaps be one varve.

7601–7750.—Although the varve limits frequently are difficult to distinguish, the two measurements correspond fairly well. The clay is fairly greasy in the lower part but grows lean upwards. The increase in thickness of the varves beginning with varve 7719 may be due to shallowing by deposition. Between varves 7613 and 7614 one varve has perhaps been skipped.

7619 and 7620 might be one varve.

7635 and 7702 contain sand and may be abnormally thick.

7658 and 7659 also may be one.

7670 should perhaps be two varves.

7726 and 7727 may be one.

*Winooski Valley*, "20 Vt.," "21 Vt.," localities 148–152. Plate V.

This series may be correctly connected with the varve series measured in the Connecticut Valley (Antevs, 1922, Pl. V), though the correspondence partly is poor. The poor agreement may be due to local conditions and then primarily to increase of the drainage areas as new tributary valleys became uncovered. As there are only two series to compare, it cannot be determined which version indicates the actual ice melting, when disagreement exists.

The different measurements in the Winooski Valley show good to excellent correspondence with each other. Those at localities 150, 151, and 152 have not been used in calculating the normal curve simply because the varves are much thicker than at localities 148 and 149.[1]

7201 and 7202 may be one varve.

7226 and 7227 may also be one.

7254 should perhaps be two varves.

7294, at localities 150 and 151, begins a series of very thick varves largely of stiff clay (cf. p. 192). The lowest varves are disturbed but are probably each a few feet thick. They evidently mark a drainage. As the material is coarser at locality 151 than at 150—the altitudes are about the same—and since there may be no other drainage, it must have been the coalescing of the lake in the Lamoille Valley with that in the Winooski Valley via the valleys of Joes Brook and Waterbury River (cf. p. 94; see also Merwin, 1908, Pl. 21). The unusually thick superjacent varves of greasy clay show that enormous quantities of fine mud were transported across the watershed.

*Essex*, Vt., localities 153–159, 164–170. Plate V.

This series, consisting of clay measurements at Essex and Cambridge, is not correlated with those in the upper Winooski Valley nor with those in the Connecticut Valley. As the watershed between Huntington River and Hollow Brook, 16 miles

[1] Varves 7059 to 7288 are the same as De Geer's (1926, Pl. 2) varves Waterbury −1418 to −1183.

*Essex—Continued*

southeast of Burlington, was first uncovered long after year 7330, the Essex region perhaps became ice-free about year 7500.

1–56.—The only curve is based on measurements made near each other. The lowest varves, deposited close to the ice edge, are thick and uncharacteristic. Varves 10–25 consist of fairly fat clay, varves 26–56 of quicksand, but with the exception of 26 and 27 they are not abnormally thick.

57–77.—The two series agree fairly well, though the varves decrease gradually in thickness from the mouth of Alder Brook on the Winooski towards Essex Center. The varves do so throughout the series, suggesting that relatively little material was brought down from the area of Browns River northeast of Essex Center, while the Winooski River after the coalescing of Lake Winooski with Glacial Lake Champlain may have begun to erode and transport the glacial deposits downstream, adding to the material coming directly from the melting ice.

57 may be abnormally thick. It may mark a drainage far away, since its coarsest material is clay silt.

78–200.—Curve A: The different measurements, made near each other, show excellent correspondence. Curve C: From varve 131, where the series measured at locality 166 begins, the agreement is for the most part good. Series A and B agree fairly well; the curve of section 155 takes an intermediate position and makes the correlation sure. Curve C shows satisfactory correspondence to the other two, especially to series B, although the varves are twice as thick as in the latter.

85, series C, is perhaps abnormally thick.

98, series B, may be too thick.

143–180, locality 168, consist largely of sandy silt or of quicksand, so that the thicknesses are mostly worthless. Most of the varves are much thicker than at

78–200.—*Continued*

locality 166, 2½ miles to the west. Only varve 180 is thicker and contains coarser material at locality 166. If localities 169 and 170 are correctly connected, the varves are several times thicker at Jeffersonville than at Cambridge. Varve 181 at localities 166 and 168 abruptly begins a stiff blue clay with varves less than half as thick as those below varve 181. At Essex no change in the deposition is noticeable. A tributary from the east or northeast that discharged large quantities of material into the Lamoille Valley must have become diverted.

143, 144, 145, series C, were measured as one varve.

156, locality 153, is 2 feet thick and consists of quicksilt with fragments of varved clay; two subjacent varves are torn up. The winter layer is only 1/6 inch thick. At locality 155 the varve is 1⅞ inch thick. It evidently marks a drainage into the Winooski Valley near locality 153. How the drainage took place is not evident; but perhaps a moraine barrier or a barrier of buried ice gave way.

164 and 165, locality 166, were measured as one varve.

180, at locality 166, consists largely of sand and at locality 168 of coarse silt. It is here abnormally thick.

184 and 187, localities 166 and 168, were each measured as two varves.

193 and 194, localities 166 and 168, were measured as one varve.

200–284.—Series A: The different measurements correspond well. Series C: The two measurements, as far as number 168 reaches, also agree well. Series A and B show fairly good correspondence, considering the great difference in the thickness of the varves. A large percentage of the material at locality 153 may have been brought down by the Winooski eroding the glacial lake beds farther up the valley. Series C agrees rather well in

200–284.—*Continued*

part, in part not well with the other two. Especially noteworthy are the thin varves 236–239 and 246 and 247 in series C.

220 contains silt and is very thick in all the series. It perhaps records a year of unusual melting.

285–345.—The different measurements in series A show good agreement. The two series correspond, though curve A is very flat. The whole series, except varve 302, consists of lean silt or sandy silt.

302 at all localities consists for the most part of stiff clay. It apparently marks a drainage that took place so far away from these sections that only fine clay was brought to them. The thickness of the stiff clay is somewhat greater at Essex Center than near Cambridge.

303–about 316, locality 166, are sandy and covered by silty sand.

303–345, series A, consist of quicksilt. The material increases in coarseness up to the ground surface. The increase in the thickness of the varves and in the coarseness of the material has probably no relation to the drainage recorded by varve 302, for the change is most marked at locality 166, where the drainage made itself least felt. The change seems to be best accounted for by a lowering of the level of the Glacial Lake Champlain, but it is not evident how that could take place at this time, when the lake discharged southward, and moreover the lake has been traced at high levels farther north (Fairchild, 1916, Pl. II). However, the events are in reality little known, so that these objections perhaps lack validity. The change might possibly also be due to the discharge of a large tributary into the Lamoille Valley. However, this is not probable, since in that case fine mud, if anything—the cessation of discharge at Cambridge year 180, as

285–345.—*Continued*

stated, did not affect the Essex region—would have been transported down to Essex Center, whereas the material here is very lean quicksilt.

*Fairfax*, Vt., localities 161–163. Plate VI.

This section is situated only about five miles west of that in the Cambridge region. The sudden increase in the thickness of the varves that takes place at varve 139 recalls the increase at varve 302 in the Essex-Cambridge series. The level of the zone of change is about 400 feet at locality 161, 450 feet at locality 163, 460 feet at Essex Center, and probably some 500 feet at Cambridge. Since, however, there is no further agreement between the series, the one at Fairfax may date from a later time. The three measurements agree well.

12 contains fine sand and marks a drainage.

75–80 contain small white lime concretions, so that they appear spotted.

80–89 consist largely of fine-grained material, which comes perhaps partly from a distant drainage.

139 begins an abrupt increase in deposition which is much more marked at locality 161, south of Fairfax, than at locality 163. The excessive material may have come from the east, from the upper Lamoille. It may have originated either where a new tributary discharged into the Lamoille Basin or where the upper Lamoille eroded the glacifluvial beds after lowering the lake level (cf. p. 94). The clay gradually becomes lean upwards and changes into silt, sandy silt, and sand.

*Locality 160*, Vt., Plate VI.

This varve series is not correlated with any other. The clay is exceedingly stiff; the winter layers mostly comprise from three-fourths to nearly the whole of the varves.

53 contains some sand at the bottom but consists otherwise of stiff clay. It marks a distant drainage.

*Locality 160—Continued*

102 consists largely of silt and indicates a drainage in the vicinity.

166–170 contains small white lime concretions.

*Orleans*, Vt., localities 175 and 176. Plates V and VI.

This series is not connected with any other. The two measurements agree perfectly with each other. The gradual upward decrease in the thickness of the varves is due to growing distance of the ice border. The great annual deposition is not necessarily an indication of rapid recession, since the basin of sedimentation was small in comparison to the drainage area. Varve 104, 7 inches (18 cm.) thick and containing sandy silt, marks a drainage in the vicinity.

*Coventry*, Vt., localities 177–179. Plate VI.

The measurements at localities 177 and 178, which lie only a mile apart, show excellent correspondence with each other. The series may be correctly connected with the upper clay at locality 179, as indicated in Plate VI. At locality 179 varve 34 was measured as two. The clay at the latter place is covered by a till bed that no doubt marks a readvance of the ice. This occurred after year 56. If this connection is correct, which is likely, the Coventry series cannot have anything in common with that obtained at localities 180 and 181, though the varves Coventry 59–79 and 84–110 correspond well with the varves Newport 1–21 and 26–53. A correlation would mean that the recession between localities 177 and 181, 8½ miles (13.7 km.) apart, represented somewhat more than 60, probably about 75 years, and that it amounted to about 600 feet (183 m.) annually. However, so rapid a retreat is improbable both because of the overriding of the clay at locality 179 and because of the moraine at Newport.

4 and 5 mark a large drainage, at some distance from locality 178. Varve 4 consists of 22 inches (55 cm.) sandy silt. Varve 5 consists of 20 inches (50 cm.) sandy silt, 10 inches (25 cm.) silt, and 14 inches (35 cm.) stiff clay.

*Coventry—Continued*

22 and 26 also contain sandy silt and mark drainages.

34 also is abnormally thick. At locality 179 it was measured as 2 varves.

68–70 contain silt and mark drainages. Varve 70 should perhaps be 2 varves.

81 is somewhat slidden. It consists largely of stiff clay. It may represent 2 varves.

*Newport*, Vt., localities 180, 181. Plate VI.

The series is not correlated with any other, strangely enough not even with that obtained just across the International Boundary (cf. p.125). The different measurements show excellent correspondence with each other.

1–12 as deposited close to the ice edge consist largely of sandy silt.

10–12 record a drainage in the neighborhood.

13–21 consist of fine silt.

The gap, the disturbed varves, contain sand.

57, 58, and 70 contain silt and may be abnormally thick.

## Southern Quebec and Ontario

The measurements at localities 144–165, Canada, are mostly not connected with each other, so that the valuation given under the description of the localities on pages 214–222 may be sufficient.

Plates VI and VII.

## Northern Ontario and Quebec

*Timiskaming*, localities 105–135. Plate VIII.

Curves, partly comprising the same varves as here given but based on measurements in other regions, were published in a previous paper (Antevs, 1925b, pp. 126–128, Plates VI, VII, VIII). In the following paragraphs comparison is made between all the curves.

*Timiskaming—Continued*

873–913.—Series B does not agree well with series A and C, but the two latter series were obtained on Lake Timiskaming some 80 miles to the southwest of the former one.

877 and 884, series B, are too thick.

914–1000.—Series B: The individual measurements made close to each other show detailed agreement. Series F: The different measurements correspond very well. Curves B and F, obtained 47 miles apart, agree rather well. They show tolerable correspondence with series A and C and thus record fairly well the relative ice melting during the different years.

1001–1200.—Series B: The measurements agree perfectly. Series F: The measurements show good to very good correspondence. Series G: The measurements agree well. These three series, with single exceptions, correspond remarkably well with each other and fairly well with the series A and C from the region of Lake Timiskaming.

1201–1400.—Series F: The individual measurements show nearly detailed correspondence. Series F and G agree tolerably well. Series F agrees fairly well with series A, D, and E. Series G corresponds fairly well with series E. Considering the distant and different location of the measurements the correspondence is remarkable.

1401–1600.—The measurements within each series, including those not used in construction of the curves, show detailed agreement with each other. Series F does not agree with the others, but, considering that it consists of clay deposited far from the ice border, it perhaps records the ice melting well. Series D and E agree excellently. Their agreement with series A, B, and C before 1528 is poor, but from 1528 on it is good.

The sudden increase in the thickness of the varves beginning with varve 1528 occurs on the Frederick House River and Black River as well as at La Sarre, east of Lake Abitibi. It is most marked in series E

1401–1600.—*Continued*

and least marked in series D, consisting of measurements made two miles farther north. The actual thickness of the varves before year 1528 is greater in series D than in series E, but from year 1528 it is greater in series E than in series D. The increase cannot be due to the more rapid settling of the mud, coming down before long transportation because of flocculation, for the character of the clay does not undergo any change. For the same reason it cannot be attributed to decreased water depth and subsequent shore erosion. The increase was believed by this writer (1925b, pp. 15, 82) to be due to augmented ice melting, but perhaps it was rather caused by increase of the drainage area. Why the increase is greater at localities 123–129 than either north or south of them is difficult to understand.

1601–1800.—Series D: The individual measurements up to 1750 present excellent correspondence. Thereafter the varves grow too thin to be characteristic. Till year 1640 the series also shows good agreement with series A and B; after that partly good, partly poor. The greater thickness of the varves in series A (localities 91 and 92), situated farther from the ice edge than series D, is remarkable. So is also the 18 inch (45 cm.) thick drainage varve year 711 at locality 92, not recorded at all in the profiles on the Frederick House River.

1801–1880.—The measurements correspond rather well. The curve is good and agrees fairly well with curve A.

1881–2027.—The varves are too thin accurately to record the ice melting, but the parts used in the construction of the curve agree well with each other. The large deposition during the years 1834–1856 may be due to relatively large ice melting. The increase beginning year 2019 at localities 124 and 125 may be due to approach of the readvancing ice edge.

*Localities 136 and 137.* Plate VII.

The two profiles agree fairly well with each other and with the Timiskaming series, and the connections indicated by the numbers given the varves may be correct.

*Locality 138.* Plates VIII and VII.

The varve series is not connected with any other (cf. p. 150). The gap above varve 24 may not be great; it is even possible that the two parts overlap and that varve 26 is the same as varve 14. The varvity is good, and the varve thicknesses may faithfully record the relative summer ice melting. In year 236 the clay changes from exceedingly stiff to moderately greasy, but the top varves are again very stiff. The quick increase in the thickness of the varves beginning with year 236 may be due to approach of the readvancing ice border. From varve 243 and upward the clay contains pieces of light-colored clay, silt, etc. Varves 184 and 185 may be one. Varve 186 may be two varves.

For localities 139–143 see the description of the localities, pages 213, 214. Plate VII.

### Toronto and Scarboro Heights

*Toronto,* localities 166–170. Plate IX. Cf. pp. 129–134.
Series A: locality 166, Don Valley.
Series B: locality 167, Don Valley.
Series C: Scarboro Bluffs.

1–200.—The summer layers consist of coarse clay or fine clay, and the winter layers, which form from one-half to two-thirds of the varves, are, as always, of extremely fine-grained clay. The thin varves are quite dark-colored, and their limits are difficult to distinguish. In the summer layers, but not in the winter layers, throughout the series are scattered stones, evidently dropped from drifting icebergs. The thin varves 94–135, both in Don Valley and at Scarboro, are so rich in fragments of white limestone and small pebbles that the zone is

1–200.—*Continued*

distinctly spotted. The curves may for the most part faithfully record the relative annual amount of the ice melting, for on the whole they show good correspondence with each other.[1]

201–400.—In the Don Valley the thin varves consist of dark-gray silty clay and fat, very thin clay layers. The summer layers of the thick varves consist of alternating layers of sandy silt and coarse clay. The winter layers are extremely thin. They bear no relation whatever to the summer deposit, so that the bulk of the fine material must have been transported farther away from the ice border.

At Scarboro the summer layers consist of clay silt which is mostly lead-colored. The winter layers, which in thin varves are moderately thick and in thick varves relatively thin, are usually brown to reddish-brown in color. Varves 338–353 contain white fragments of limestone and are spotted. The abnormally thick varves contain silt, or, as varve 365, some sand.

The two curves from Don Valley on the whole correspond fairly well, but these do not agree well with the curve from the Scarboro Heights. The most marked difference is the often incomparably greater thickness of the varves in Don Valley. The alternation of groups of thick varves with groups of thin varves most probably was due to shifting of the outlet of the glacier river or swings of the transporting current. Decreasing water depth may have been at least a contributing factor to the increase of the varve thicknesses in the uppermost part of section 166. It is less probable that the other great variations were due to changes in the lake level.

401–586.—The summer layers consist of silty clay and silt, and in

[1] Varves 1 to 234 are the same as De Geer's (1926, Pl. 2) varves Toronto −1420 to −1173.

401–586.—*Continued*

the thick varves also of fine sand. The winter layers are so thin that the varve limits in some cases are difficult to determine. The differences in thickness are frequently too great to be due to the variation in the annual ice melting. The thickest varves must be caused by drainages of ice-dammed lakes, swings of the transporting current, etc.

*Locality 174.* Plate IX.

Bed 12 in section 174, of which the curve presented is a part, is younger than the other clays at Scarboro of which curves are reproduced. As the lowest part of the bed is somewhat disturbed, the series given begins about 25 varves above the sand. The clay is rather stiff, as the winter layers represent from one-third to one-half of the varves. The varvity is very distinct. Varve 170 begins a pressed series, 9 inches thick, of dark, spotted clay with thin varves—about 40 in number. This series is overlain by a foot of clay with 27 distinct varves forming the uppermost part of bed 12.

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# CONVERSION TABLES

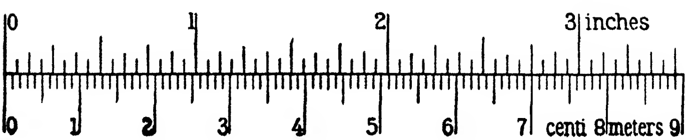
| | Inches to Centimeters | Feet to Meters | Miles to Kilometers | Centimeters to Inches | Meters to Feet | Kilometers to Miles |
|---|---|---|---|---|---|---|
| 1 = | 2.54000 | 0.304800 | 1.60935 | 0 39370 | 3.28083 | 0.62137 |
| 2 = | 5.080 | 0.609601 | 3.21869 | 0.78740 | 6.56167 | 1.24274 |
| 3 = | 7.620 | 0.914402 | 4.82804 | 1.18110 | 9 84250 | 1.86411 |
| 4 = | 10.160 | 1.219202 | 6.43739 | 1.57480 | 13.12333 | 2.48548 |
| 5 = | 12.700 | 1.524003 | 8 04674 | 1 96850 | 16 40417 | 3.10685 |
| 6 = | 15.240 | 1.828804 | 9 65608 | 2 36220 | 19 68500 | 3.72822 |
| 7 = | 17.780 | 2.133604 | 11.26543 | 2.75590 | 22.96583 | 4.34959 |
| 8 = | 20.320 | 2.438405 | 12.87478 | 3 14960 | 26 24667 | 4.97096 |
| 9 = | 22.860 | 2.743205 | 14.48412 | 3.54330 | 29 52750 | 5.59233 |
| 10 = | 25.400 | | | | | |
| 11 = | 27.940 | | | | | |
| 12 = | 30.480 | | | | | |

1 foot = 12 inches.
1 yard = 3 feet.
1 fathom = 6 feet.
1 mile = 5280 feet = 1760 yards.
1 square mile = 2.589998 square kilometers.
1 cubic mile = 4.16819 cubic kilometers.

1 meter = 1.093611 yards.
1 square kilometer = 0.3861006 square mile.
1 cubic kilometer = 0.239912 cubic mile.

$F° = \frac{9}{5}C° + 32°$.
1 F° = 0.56 C°.

$C° = \frac{5}{9}(F° - 32°)$.
1 C.° = 1.8 F.°



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